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# **Reconstructing ice dynamics in the central sector of the last British- Irish Ice Sheet**

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**A thesis submitted in partial fulfilment of the requirements of the  
University of Durham for the degree of Doctor of Philosophy.**

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University of Durham**

## Abstract

The central sector (NW England and Scottish borders) of the last British-Irish Ice Sheet exhibits a palimpsest glacial geological and geomorphological signature characteristic of multi-phase ice flow and ice-marginal fluctuations. Despite its influential position at the heart of the British-Irish Ice Sheet, sourced from major ice dispersal centres of the northern Pennines, Lake District and Southern Uplands, and drained via fast-flowing outlets such as the Irish Sea Ice Stream, the region remains poorly constrained, both temporally and in terms of ice-flow dynamics. The principle goal of this thesis was therefore to reconstruct the palaeoglaciology of the central sector of the British-Irish Ice Sheet during the last glacial cycle, focusing on: (1) ice-flow dynamics with respect to palaeo-ice divides, ice-dispersal centres, flow trajectories and flow phasing; (2) the relative chronology of ice flows during advance and decay of the ice sheet; and (3) evidence for ice stream activity either within or sourced from the study area.

The thesis adopted a dual approach involving both geomorphological mapping and sedimentological analysis. A 5 m resolution NEXTMap DEM was used to map over 9,000 individual landforms including subglacial lineations, hummocky terrain, moraines, meltwater channels, eskers and glaciofluvial sediment accumulations. Subglacial lineations were subdivided into discrete flow sets demarcating distinctive flow phases, and a relative chronology produced from cross-cutting relationships. Thirteen field sites, concentrated in the Solway Lowlands, supported by data collected from over 200 boreholes enabled detailed stratigraphic and sedimentological analysis to be carried out. This included stratigraphic logging, the collection of macrofabrics, particle size and geochemistry analysis on till matrixes, clast lithological counts, varve analysis and microstructural (thin sections) data.

Results from this study have demonstrated that the central sector of the ice sheet was characterised by repeated ice-flow switches, initiation and termination of ice streams, drawdown into ice streams, repeated ice-marginal fluctuations (the Scottish and Blackhall Wood Re-advances) and the production of large volumes of meltwater, often impounded to form ice-dammed lakes. Six main stages of ice flow have been recognised in the region, of which stage I is thought to indicate the period of maximum ice expansion, while stages II-VI record the deglacial history. A pre-stage I event is also discussed and can be reconciled with the initial expansion of ice out of upland dispersal centres. Stage I was characterised by ice flowing eastwards across the country through major topographic lows of the Stainmore and Tyne gaps. The Tyne Gap was occupied by a topographic ice stream, which was heavily influenced by the changing dominances of both Lake District and Southern Upland ice-dispersal centres. Migration of ice divides back towards upland dispersal centres during stage II resulted in the flow of ice through the Stainmore Gap being cut-off, while the northern edge of the Tyne Gap ice stream was breached by a SE ice flow down the N Tyne Valley. Despite the maintenance of the Irish Sea Ice Stream off the western coast of Cumbria throughout stage III, the Tyne Gap and Solway Lowlands underwent widespread deglaciation. Meltwater from the Tyne Gap was routed into Glacial-Lake Wear via a major proglacial drainage

network in the South Tyne Valley, while the natural basin of the Solway Lowlands also ponded-up (Glacial-Lake Blackhall Wood) as drainage became impeded by the Irish Sea Ice Stream. The overall pattern of retreat was reversed during the Blackhall Wood Re-advance (stage IV), during which ice was vigorously drawn down into the Irish Sea Ice Stream. Stage V was characterised by the continued retreat of ice out of the central sector of the British-Irish Ice Sheet; with the vast amounts of meltwater generated impounded in ice-marginal lake systems (Glacial-Lake Carlisle), or routed through meltwater channel networks or evolving glacier karst (Brampton kame belt). The landforms of the Brampton kame belt can be reconciled with ice stagnation on the reverse slope of the Tyne Gap, and is thought to have formed one component of a much larger, time-transgressive drainage network involving the Pennine escarpment and Tyne Gap meltwater channel systems. The final recognised stage in the glacial history of the region was the Scottish Re-advance, a brief incursion of ice, sourced from the Southern Uplands, onto the fringe of the Solway Lowlands. A large glacial lake is identified to have formed at the ice front, dammed against ice in the Irish Sea basin and delimited by a large deltaic complex at Holme St. Cuthbert.

## Contents:

List of Tables	xii
List of Figures	xiv
Acknowledgments	xxi
Declaration	xxiii
<b>CHAPTER 1: Introduction</b>	<b>1</b>
1.1 Rationale	1
1.2 Motivation and research questions	3
1.3 Region of interest	4
1.3.1 <i>Topography, geology and geomorphology</i>	4
1.3.2 <i>General palaeoglaciological setting</i>	8
1.4 Overview of materials and methods	9
1.4.1 <i>Glacial geomorphological mapping</i>	9
1.4.2 <i>Glacial stratigraphy, sedimentology and provenance</i>	12
1.5 Overview of results	18
1.5.1 <i>Chapter 2</i>	18
1.5.2 <i>Chapter 3</i>	18
1.5.3 <i>Chapter 4</i>	19
1.5.4 <i>Chapter 5</i>	19
1.5.5 <i>Chapter 6</i>	20
1.5.6 <i>Chapter 7</i>	21
1.5.7 <i>Chapter 8</i>	21
1.6 References	22

<b>CHAPTER 2: Glacial geomorphology of the central sector of the last British-Irish Ice Sheet</b>	<b>32</b>
2.1 Introduction	32
2.2 Methods	33
2.2.1 <i>Dataset</i>	33
2.2.2 <i>Mapped landforms</i>	34
2.2.3 <i>Identifying complex flow</i>	41
2.2.4 <i>Elongation ratios</i>	44
2.3 Implications for BIIS flow dynamics	44
2.4 Conclusions	48
2.5 Software	48
2.6 Acknowledgments	48
2.7 References	49
 <b>CHAPTER 3: The palaeoglaciology of the central sector of the British and Irish Ice Sheet: reconciling glacial geomorphology and preliminary ice sheet modelling</b>	 <b>53</b>
3.1 Introduction	53
3.2 Previous palaeoglaciological reconstructions	54
3.3 Rationale and approach	61
3.4 Glacial geomorphology	64
3.4.1 <i>Eastward flowing ice in the Tyne Gap</i>	65
3.4.2 <i>Opposed ice flow patterns in the Solway Lowlands</i>	67
3.4.3 <i>Bedform overprinting in the Vale of Eden</i>	69
3.4.4 <i>Late stage unconstrained flows in the Solway Firth</i>	69
3.4.5 <i>Regional summary of major ice flow phases</i>	72
3.5 Ice flow reconstruction from numerical modelling	73

3.5.1	<i>Model description</i>	73
3.5.2	<i>Settings and tuning of model parameters</i>	74
3.5.3	<i>Results and discussion of modelling</i>	76
3.6	Integration of modelling and glacial geomorphology	80
3.7	Conclusions	84
3.8	Acknowledgments	85
3.9	References	85
 <b>CHAPTER 4: A major ice drainage pathway of the last British-Irish Ice Sheet: the Tyne Gap, northern England</b>		94
4.1	Introduction	94
4.2	Study area	96
4.3	Previous research on the Late Devensian glaciation of the Tyne Gap	98
4.3.1	<i>Subglacial bedform evidence for ice flow</i>	98
4.3.2	<i>Erratics</i>	98
4.3.3	<i>Till stratigraphy in northern England</i>	98
4.3.4	<i>Deglaciation</i>	99
4.4	Methods	99
4.4.1	<i>Subglacial bedform mapping</i>	100
4.4.2	<i>Sedimentology and stratigraphy</i>	100
4.5	Results and Interpretation	101
4.5.1	<i>Glacial Geomorphology</i>	101
4.5.1.1	<i>Streamlined subglacial bedforms</i>	101
4.5.1.2	<i>Relative chronology of flow sets</i>	106
4.5.1.3	<i>Meltwater channels</i>	107
4.5.1.4	<i>Moraines, glaciofluvial and glaciolacustrine deposits</i>	108

4.5.2	<i>Sedimentology and stratigraphy</i>	109
4.5.2.1	<i>Willowford</i>	109
4.5.2.2	<i>Hayden Bridge</i>	117
4.5.2.3	<i>Borehole logs</i>	119
4.6	Discussion	122
4.6.1	<i>Ice sheet dynamics through the Tyne Gap corridor</i>	122
4.6.2	<i>A palaeo-ice stream?</i>	127
4.6.3	<i>Wider implications for the British-Irish Ice Sheet</i>	128
4.7	Conclusions	130
4.8	Acknowledgments	130
4.9	References	131
 <b>CHAPTER 5: The Brampton kame belt and Pennine Escarpment meltwater channel system (Cumbria, UK): Morphology, sedimentology and formation</b>		139
5.1	Introduction	140
5.2	Process-form models of kame production and previous work on the Brampton area	140
5.3	Methods	145
5.3.1	<i>Geomorphological mapping</i>	145
5.3.2	<i>Sedimentology and stratigraphy</i>	146
5.4	Geomorphology	146
5.5	Sedimentology and stratigraphy	153
5.5.1	<i>Ridges</i>	153
5.5.1.1	<i>Kirkhouse</i>	153
5.5.1.2	<i>Brampton Ridge</i>	159
5.5.1.3	<i>Braithwaite's sand and gravel pit</i>	160

5.5.1.4	<i>Boreholes</i>	161
5.5.2	<i>Flat-topped hills</i>	163
5.5.2.1	<i>Faugh sand pit</i>	163
5.5.2.2	<i>Whin Hill, Faugh sand and gravel pit</i>	170
5.5.2.3	<i>Borehole data</i>	170
5.6	Discussion	171
5.6.1	<i>Geomorphology and sedimentology</i>	171
5.6.2	<i>Model of formation for the Brampton kame belt</i>	176
5.7	Conclusions	180
5.8	Acknowledgments	181
5.9	References	181
<b>CHAPTER 6: Sedimentary evidence for a major glacial oscillation and pro-glacial lake formation in the Solway Lowlands (Cumbria, UK) during Late Devensian deglaciation</b>		190
6.1	Introduction	190
6.2	Study area and glacial geomorphology	194
6.3	Methods	194
6.4	Sedimentology and stratigraphy	195
6.4.1	<i>Blackhall Wood</i>	195
6.4.2	<i>Maryport</i>	210
6.4.3	<i>Swarthy Hill</i>	213
6.4.4	<i>Borehole stratigraphy</i>	215
6.5	Discussion	219
6.6	Conclusions	225
6.7	Acknowledgments	226



6.8	References	226
<b>CHAPTER 7: Re-advance of Scottish ice into the Solway Lowlands (Cumbria, UK) during the Main Late Devensian deglaciation</b>		236
7.1	Introduction	236
7.2	Previous research on Scottish Re-advance	237
7.3	Methods	241
7.3.1	<i>Glacial geomorphological mapping</i>	241
7.3.2	<i>Sedimentology and stratigraphy</i>	242
7.4	Results and Interpretation	242
7.4.1	<i>Holme St. Cuthbert sand and gravel complex</i>	242
7.4.2	<i>Regional stratigraphy</i>	253
7.4.3	<i>Regional glacial geomorphology</i>	259
7.5	Discussion	262
7.5.1	<i>What geomorphological and sedimentological evidence can be correlated to the re-advance of Scottish ice?</i>	262
7.5.2	<i>What was the maximum extent of the Scottish Re-advance?</i>	264
7.5.3	<i>Was there was a concurrent re-advance of ice from the Lake District?</i>	266
7.5.4	<i>What are the glaciological implications of the Scottish Re-advance?</i>	266
7.6	Conceptual model of the Scottish Re-advance and its wider implications	267
7.7	Conclusions	270
7.8	Acknowledgments	271
7.9	References	271
<b>CHAPTER 8: The application of geochemistry and particle size analysis to the investigation of glacial stratigraphy in the central sector of the last British-Irish Ice Sheet</b>		281

8.1	Introduction	281
8.2	Physiographic setting	283
8.2.1	<i>Bedrock geology</i>	283
8.2.2	<i>Quaternary stratigraphy</i>	283
8.3	Methods	287
8.3.1	<i>Sample preparation and analysis</i>	287
8.3.2	<i>Statistical analysis</i>	289
8.4	Results	289
8.4.1	<i>Geochemical analysis</i>	289
8.4.2	<i>Particle size analysis</i>	295
8.5	Discussion	297
8.5.1	<i>Vertical stratigraphies</i>	298
8.5.2	<i>Stratigraphic framework for the Solway Lowlands-Tyne Gap</i>	299
8.6	Conclusions	302
8.6.1	<i>Statistical approach to stratigraphic correlations</i>	302
8.6.2	<i>Ice-flow history of the Solway Lowlands-Tyne Gap</i>	302
8.7	References	303
<b>CHAPTER 9: Glacial history of the central sector of the last British-Irish Ice Sheet</b>		308
9.1	Introduction	308
9.2	Overview of the glacial history	310
9.2.1	<i>Pre-stage I</i>	310
9.2.2	<i>Stage I</i>	311
9.2.3	<i>Stage II</i>	314
9.2.3.1	<i>Stainmore Gap</i>	314

9.2.3.2	<i>Tyne Gap</i>	316
9.2.4	<i>Stage III</i>	316
9.2.5	<i>Stage IV</i>	318
9.2.6	<i>Stage V</i>	321
9.2.7	<i>Stage VI</i>	325
9.3	Conclusions	327
9.4	References	328

## List of Tables

### CHAPTER 1:

1.1	Laboratory errors ( $\pm$ mg/kg) associated with ICP-MS Total Metal Extraction technique.	17
-----	---	----

### CHAPTER 2:

2.1	Mapped glacial landforms.	34
2.2	Elongation ratios: statistics.	45

### CHAPTER 3:

3.1	Mapped glacial landforms.	64
-----	---------------------------	----

### CHAPTER 4:

4.1	Statistical summary of elongation ratios for ice flow phases 2-6 in the Tyne Gap (phase 1 not in Tyne Gap).	107
-----	---	-----

### CHAPTER 5:

5.1	Whin Hill, Faugh (Huddart, 1970).	170
-----	-----------------------------------	-----

### CHAPTER 6:

6.1	Simple Event Stratigraphy for the central sector of the British-Irish Ice Sheet.	193
6.2	Average clast lithology (%) from Blackhall Wood, Maryport and Swarthy Hill, 8-64 mm.	201
6.3	Updated Event Stratigraphy and regional correlations for the glacial history of the Solway Lowlands.	223

## CHAPTER 7:

7.1	Ripple structures.	247
-----	--------------------	-----

## CHAPTER 8:

8.1	Simple Event Stratigraphy for the Solway Lowlands, showing BGS Lithostratigraphic Formations and glacial geomorphological flow-sets.	285
8.2	Principle Components.	291
8.3	Particle size analysis: percentage range of the sand, silt and clay fractions for each field site.	295
8.4	Updated Event Stratigraphy which illustrates the stratigraphic framework, provenance, related flow-phases and the evidence for assigning geochemical samples to particular flow-phases and source regions.	297

## CHAPTER 9:

9.1	Event Stratigraphy for the central sector of the last British-Irish Ice Sheet.	309
-----	--	-----

## List of Figures

### CHAPTER 1:

1.1	Map of the central sector of the BIIS, including topography and places of interest.	5
1.2	(a) 1:625k bedrock geology map of northern England and the Scottish Borders; (b) generalised geological section running west to east across the region (reproduced from King, 1976); and (c) structural elements in northern England (reproduced from King, 1976).	6
1.3	(a-c) Effect of azimuth bias on morphological analysis of glacial lineaments in the Caldew Valley.	11
1.4	Microstructures identified within thin sections of glacial sediments (van der Meer, 1993).	15

### CHAPTER 2:

2.1	Location and topography of the study area.	33
2.2	Examples of glacial landforms.	36
2.3	Examples of complex flow in the central sector of the BIIS.	43
2.4	Ice streaming round the northern Lake District into the Irish Sea Basin.	46
2.5	Generalised flow phases of ice dynamics in the central sector of the last BIIS during the Late Devensian.	47

### CHAPTER 3:

3.1	Location map of the central sector of the BIIS with areas covered by case studies located by boxes and names of sites mentioned in text.	55
3.2	Early interpretations of glacier flow directions in the central BIIS based on Trotter (1929, above) and Hollingworth (1931, below).	56
3.3	Map of cross-cutting flow sets in SW Scotland (from Salt & Evans 2004).	60

3.4	Schematic diagram to show the procedures employed in the grouping of subglacial flow sets (from Clark, 1997).	62
3.5	Three conceptual models of subglacial bedform overprinting (from Clark 1997).	63
3.6	Glacial geomorphic evidence of eastward flowing ice in the Tyne Gap.	66
3.7	Geological influence of bedrock structure on subglacial lineations.	67
3.8	Glacial geomorphic evidence of opposed ice-flow pattern in the Solway Lowlands.	68
3.9	Glacial geomorphic evidence of bedform overprinting in the Eden Valley.	70
3.10	Glacial geomorphic evidence of late stage unconstrained flows in the Solway Firth.	71
3.11	Summary maps of major flow events identified in the subglacial bedform mapping.	73
3.12	(a) Temperature anomaly from present with time derived from $\delta^{18}\text{O}$ ice core record (Daansgard et al 1993; Huybrechts 1997); (b) modelled ice sheet volume of the BIIS over the last 45 ka.	75
3.13	(a-d) Maps of modelled surface elevation and ice flow trajectories of the BIIS during build up to the LGM.	76
3.14	(a-i) Maps of modelled ice surface elevation of the BIIS during retreat from the LGM.	77
3.15	(a-d) Maps of modelled surface elevation and ice flow trajectories during deglaciation of the BIIS.	79
3.16	Map of overlain ice flow trajectories of the 4 flowsets for the different times slices in Figure 3.14.	80

#### CHAPTER 4:

4.1	(a) Location map of the study area, (b) Topography of the Tyne Gap and the surrounding area, (c) The Tyne Gap: topography, place names, major rivers and field locations.	95
4.2	1:625k bedrock geology map of northern England and the Scottish Borders.	97

4.3	(a) NEXTMap DEM of the heavily lineated Tyne Gap; (b) mapped subglacial bedforms; (c) bedrock structure: and (d) flow phases identified by cross-cutting relationships.	102
4.4	Composite stratigraphic log from Willowford field site (lithofacies codes from Benn and Evans, 1998).	110
4.5	Photographs (a-g) of lithofacies exposed at Willowford.	111
4.6	Thin sections (a-b) of sediments at Willowford field site.	113
4.7	(a) Clast macrofabric data for the two tills within LFA W4, at Willowford; (b) Ternary Diagram showing clast fabric shape (cf. Benn, 1994); and (c) Modality-Isotropy plot (cf. Hicock <i>et al</i> , 1996).	114
4.8	Provenance data at Willowford.	115
4.9	Borehole logs (BGS) showing stratigraphic sections and borehole locations at Hayden Bridge field site.	117
4.10	Photographs (a-d) of lithofacies exposed at Hayden Bridge	118
4.11	Borehole logs showing stratigraphic sections and borehole locations at Haltwhistle Bypass (BGS).	120
4.12	Maps (a-d) showing inferred ice flow stages (stages I-III) within the Tyne Gap.	123

## CHAPTER 5:

5.1	Topographic map showing the position of the Brampton kame belt within the central sector of the BIIS, and the location of fieldsites and borehole logs.	141
5.2	Schematic model for the various types of kames (from Brodzikowski & van Loon, 1991).	142
5.3	Retreat phenomena in east Edenside demarcating a series of ice marginal positions from the kame belt and meltwater channels (from Trotter, 1929).	144
5.4	Meltwater channels along the Pennine escarpment and within the Penrith sandstone outcrop (from Arthurton & Wadge, 1981).	145
5.5	(a) NEXTMap image of the Brampton kame belt and surrounding area, with meltwater channels, eskers and glacial sand and gravel deposits identified; (b)	147



	Glacial geomorphological map of the Brampton kame belt, flat-topped hills, ridges, meltwater channels and depressions; and (c) Glacial geomorphological	
5.6	Oblique aerial photograph (Cambridge University) looking north-west along the edge of Talkin Tarn, revealing a series of subdued ridges (marked x).	150
5.7	(a) Photograph showing a typical Type 1 meltwater channel from the Pennine escarpment; and (b) photograph showing a typical Type 3 meltwater channel which plunges down the Pennine escarpment and is deeply incised.	152
5.8	Location and stratigraphic logs: Kirkhouse sand pit.	155
5.9	Photographs (a-e) of lithofacies from Kirkhouse sand pit	157
5.10	Stratigraphic logs: Brampton ridge sand pit (reproduced from Huddart, 1970).	159
5.11	Stratigraphic logs: Braithwaite's sand pit (reproduced from Huddart, 1970).	161
5.12	Borehole logs through several ridges and flat-topped hills, reproduced from Jackson (1979).	162
5.13	Locations and stratigraphic logs: Faugh sand pit.	164
5.14	Photographs (a-f) of lithofacies from Faugh sand pit.	166
5.15	Photographs (a-f) of modern analogues from Iceland.	172
5.16	(a) Model of formation of the Brampton kame belt (based on the model by Brodzikowski & van Loon, 1991); and (b) sketch of a glacial-karst landsystem.	178
CHAPTER 6:		
6.1	Topographic map showing key locations and fieldsites.	192
6.2	Glacial geomorphology of the Solway Lowlands.	193
6.3	Blackhall Wood section logs.	196
6.4	Photographs (a-f) of lithofacies from Blackhall Wood field site.	197
6.5	Thin sections (a-d) from Blackhall Wood.	198
6.6	Clast macro-fabrics for Blackhall Wood.	200
6.7	Microphotographs (a-f) of typical laminated sediment from Blackhall Wood.	205
6.8	Varve thickness and count.	208

6.9	Maryport section logs and macro-fabrics.	210
6.10	Photographs of lithofacies from Maryport (a-c) and Swarthy Hill (d-f).	212
6.11	Swarthy Hill section logs, sketch of clastic dykes and fabrics.	214
6.12	(a) Field sites, borehole locations and stratigraphic sections from prior research (Goodchild, 1875; Dixon <i>et al.</i> , 1926; Huddart, 1970; Arthurton & Wadge, 1981); (b) contour map of the thickness of laminated clay and silt within the Vale of Eden, mapped against topography; (c) contour map of the thickness of the upper diamicton mapped against the NEXTMap DEM.	217
6.13	Stratigraphic logs from key sites through the Vale of Eden.	218
6.14	Glacial history of the ‘Blackhall Wood’ Re-advance in the Solway Lowlands (a-c).	221
CHAPTER 7:		
7.1	(a) Topographic map showing fieldsites and key locations; and (b) NEXTMap DEM showing the flowsets and glacial features.	239
7.2	Map of the various models of maximum extent for the Scottish Re-advance: (a) Trotter, (1929); and (b) Huddart, (1994).	240
7.3	NEXTMap DEM of Holme St. Cuthbert glaciofluvial complex superimposed over an arcuate set of lineations.	243
7.4	Photographs: (a) Looking SW down the ice-contact slope at Overby; (b) Subdued esker near Thursby (with sharp bends); and (c) Trotter & Hollingworth (1932): picture of delta sedimentology at Warwick Bridge delta.	244
7.5	Stratigraphic logs of Overby sand pit.	246
7.6	Photographs of lithofacies from Overby (a-c) and Aldoth (d-f) sand pits	248
7.7	Stratigraphic logs from Aldoth sand quarry.	250
7.8	Stratigraphic logs and clast macrofabric data from Plumpe Farm field site.	254
7.9	Photographs of lithofacies from Plumpe Farm (a-b) and Carleton (c-e) field sites.	255
7.10	Stratigraphic logs and clast macrofabric data from Carleton field site.	257

7.11	(a) 3D NEXTMap image of the glacial meltwater channels of Biglands, Wiza Beck and Wampool incised into an arcuate flowset trending round the northern margin of the Lake District; (b) plan-view of meltwater channels; and (c) meltwater channel long-profiles.	261
7.12	Borehole data (BGS) from, the region SW of Wigton.	263
7.13	Reconstruction of Scottish Re-advance into the Solway Lowlands.	268

## CHAPTER 8:

8.1	Location Map, including fieldsites.	282
8.2	1:625k bedrock geology map of the Solway Lowlands and Tyne Gap.	284
8.3	Regional ice flow dynamics in the central sector of the BIIS during the Main Late Devensian	286
8.4	Composite stratigraphic logs of all the field sites, plus locations and names of samples taken for geochemical and particle size analysis.	288
8.5	PCA: Plot of loading for the first two components.	290
8.6	PCA: Score plots for components 1-3: (a) component 1 vs component 2; (b) component 1 vs component 3; and (c) component 2 vs component 3.	292
8.7	HCA dendrogram: variables.	293
8.8	HCA dendograms with groupings (phenon line): (a) ward linkage; (b) average linkage; (c) complete linkage.	294
8.9	Ternary plots: particle size analysis: (a) Blackhall Wood; (b) Willowford; and (c) all sites.	296
8.10	Macrofabric data for Glasson field site.	300
8.11	Proposed stratigraphic framework for the diamictons of NW England.	301
8.12	Photograph: mosaic of the glacioteconite and upper diamicton at Bell Bridge.	302

## CHAPTER 9:

9.1	Stage I: Eastwards ice flow through prominent topographic corridors of the north Pennines.	313
-----	--	-----

9.2	Stage II: Cut-off of the Stainmore Gap as an ice through-flow and northwards migration of the Solway Firth ice divide.	315
9.3	Stage III: Westerly retreat of ice out of the Tyne Gap and the development of an ice-free enclave within the Solway Lowlands.	317
9.4	Stage IV: Blackhall Wood Re-advance.	320
9.5	Stage V: Retreat of ice out of the Solway Lowlands: (a) illustrates the initial retreat of both Scottish and Lake District ice out of the Irish Sea Basin; and (b) refers to a later phase during the same retreat stage associated with the ‘unzipping’ of Lake District and Scottish ice.	322
9.6	Stage VI: Scottish Re-advance.	326

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## **Declaration**

This thesis embodies the results of original research carried out between October 2006 and February 2010. References to existing works are made as appropriate. Any remaining errors or omissions are the responsibility of the author.

Stephen J. Livingstone

25<sup>th</sup> February, 2010

## Chapter 1

# Introduction

### 1.1 Rationale

Ice sheets are a crucial component of the climate-system and are capable of regulating, driving and responding to climate change. They are not simple systems that behave linearly to internal and external forcing but rather are a complex series of sub-systems, including ice domes, ice streams, outlet glaciers and ice-shelves, which are characterised by different lag times, sensitivities and velocity fields. Therefore, an understanding of the dynamics, internal organisation, stability and evolution of both palaeo and contemporary ice sheets is crucial to reconciling their interactions with the climate and environment (Alley *et al.*, 2005). This is especially relevant given that it is now thought that major changes in ice sheet dynamics caused by external forcings, can occur on timescales of years to decades (Bamber *et al.*, 2007). Furthermore, recent observations highlight a recognisable, episodic acceleration and thinning of both West Antarctic and Greenland ice sheet margins along outlet glaciers and ice streams (Shepherd & Wingham, 2007). This has resulted in renewed interest in the potential for the West Antarctic Ice Sheet to collapse (cf. Vaughan, 2008). An understanding of the glacial history and behaviour of former ice sheets can provide an important analogue for the changes currently occurring at contemporary ice sheets (Greenland and Antarctica), especially in regions with similarly large marine based sectors, such as the British-Irish Ice Sheet (BIIS).

The palaeo-record provides compelling evidence that during the last glaciation, ice sheets of the northern hemisphere underwent large changes in volume, periodic catastrophic discharges of icebergs (Heinrich events) and fluctuations in ice sheet geometry (Heinrich, 1988; Grousset *et al.*, 1993; Broecker, 1994; Bond & Lotti, 1995; Fronval *et al.*, 1995; Scourse *et al.*, 2000; Peck *et al.*, 2007). These changes, in response to sub-Milankovitch events (e.g. Bond *et al.*, 1992, 1993; MacAyeal, 1993a,b), provide clear analogues for future ocean-cryosphere-atmosphere sensitivities to climate change in areas such as the Arctic Ocean, Greenland and Antarctica. However, for palaeoglaciological research to significantly improve the boundary conditions for modelling studies, assessments of ice sheet growth and decay rely on accurate reconstructions of ice sheet volume and behaviour through time, and these must be informed by good quality and systematic field data.

Recent modelling studies have demonstrated that the BIIS was highly dynamic, drained by a number of oscillating, fast-flowing ice streams (Boulton & Hagdorn, 2006; Hubbard *et al.*, 2009) and associated with rapid switches in ice flow direction, driven by shifting ice dispersal centres and



ice divides (Clark & Meehan, 2001; Salt & Evans, 2004; Evans, *et al.*, 2009). These reconstructions are in accord with the field evidence, which is characterised by a rich diversity and complexity of Late Pleistocene sediments and landforms (Clark *et al.*, 2004). The BIIS is now thought to have extended as far south as the Isles of Scilly (Scourse, 1991; Hiemstra *et al.*, 2006), covered almost all of Ireland (Ó Cofaigh & Evans, 2007), extended to the northern continental shelf edge (Bradwell *et al.*, 2007, 2008) and coalesced with Scandinavian ice in the North Sea Basin (Carr, *et al.*, 2006; Bradwell *et al.*, 2008; Sejrup *et al.*, 2009). It was drained by large, fast-flowing ice streams and outlet glaciers in the Celtic Basin (Evans & Ó Cofaigh, 2003), fingering out from the north-western and north-eastern edge of Scotland (Merritt *et al.*, 1995; Bradwell *et al.*, 2007), and down the east coast of England (Boulton & Hagdorn, 2006). Despite a general convergence of opinion, the maximum extent of the BIIS in certain regions has not been completely resolved, while several other problems remain; particularly with respect to the thickness and timing. A paucity of dates has hindered attempts to constrain the temporal evolution of the ice sheet (e.g. Telfer *et al.*, in press) while ice thicknesses recorded from trimlines (Ballantyne *et al.*, 1998) have caused some controversy due to problems with equifinality arising from formation by both englacial thermal boundaries and periglacial activity (e.g. Boulton & Hagdorn, 2006; Kleman & Glasser, 2007). Deglaciation was typified by periodic calving into the Atlantic Ocean (Knutz *et al.*, 2001, 2002; Peck *et al.*, 2007), pro-glacial lake development along its southern margin (Evans *et al.*, 2005; Bateman *et al.*, 2008), oscillating margins (Evans & Ó Cofaigh, 2003; Thomas *et al.*, 2004; Thomas & Chiverrell, 2007; McCabe *et al.*, 2007) and numerous re-advances, some of which have been correlated with Heinrich events (McCabe & Clark, 1998; McCabe *et al.*, 2005).

Over the last decade the glacial landscape of the BIIS has been increasingly recognised to exhibit a signature of dynamic ice flow (e.g. Salt, 2001; Salt & Evans, 2004; Greenwood & Clark, 2008). This paradigm shift from a static model of ice dispersal (e.g. Bowen, 2002) to dynamic flow phases caused by migrating ice divides and dispersal centres was driven by conceptual breakthroughs and glacial geomorphic mapping of subglacial bedforms (Punakri, 1982, 1993; Boulton & Clark, 1990a,b; Clark, 1997; Clark *et al.*, 2000; Clark & Meehan, 2001). Indeed it is now recognised that several generations of subglacial lineations, or other glacial features can cross-cut each other, resulting in palimpsest flow signatures which can document changing ice flow directions and record a relative chronology of flow phasing (e.g. Boulton & Clark, 1990a,b; Clark, 1999; Clark & Meehan, 2001). Despite concerted recent efforts, the palaeoglaciology of the BIIS remains poorly constrained, both temporally and in terms of ice dynamics compared to other glaciated regions (e.g. Punkari, 1982; Boulton & Clark, 1990a,b).

The central sector of the BIIS is characterised by a complex record of flow phasing preserved in the glacial landform-sediment assemblages (Goodchild, 1875, 1887; Trotter, 1929; Hollingworth, 1931). However, the palaeo-glacial dynamics of the region, the relative and absolute chronology of these events, their implications for the growth and decay of the BIIS as a whole and how they relate

to North Atlantic climate change are yet to be fully reconciled; previous attempts to map the region (e.g. Trotter, 1929) resulted in a series of implausible glaciological scenarios and unanswered problems. Because of its location at the heart of the BIIS, this region holds the key to unlocking: (a) the palaeo-dynamics of the ice sheet through the last glacial cycle; and (b) the linkages between major ice flow phases, divide shifts, ice sheet marginal oscillations and sub-Milankovitch scale climate events.

## 1.2 Motivation and Research Questions

This study was motivated by the need for a detailed reconstruction of the palaeoglaciology of the former centre of the BIIS during the last glacial cycle. Comprehensive field mapping has been carried-out for discrete blocks of the field area (e.g. Trotter, 1929, Hollingworth, 1931). However, this research has failed to provide either a holistic map of the glacial geomorphology or a glaciologically plausible reconstruction which explains all available evidence. This is in part a consequence of the complex geomorphology and stratigraphy which has proved difficult to reconcile and is open to considerable debate (see Huddart, 1970 and Smith, 2002 for the most up-to-date reconstructions). However, recent developments in GIS technologies have seen the synthesis of a 5 m resolution NEXTMap DEM which is now available for the whole of Great Britain. This research therefore benefits from and is partly motivated by this new DEM which allows mapping over a wide spatial scale and at a high resolution. The field area chosen for this thesis is situated at the core of the BIIS and as such is intimately associated with ice flow over much of the ice sheet. Therefore, resolving the complexities associated with this area has implications not only in terms of my chosen study site, but also throughout the BIIS. Specific research questions are:

- i. *What was the palaeoglaciology of the central sector of the BIIS with respect to palaeo-ice divides, ice dispersal centres, flow trajectories and flow phasing?*

Despite localised mapping of various sectors of the central region of the BIIS (cf. Trotter, 1929; Hollingworth, 1931), a holistic model of regionally integrated ice flows has yet to be achieved. This is crucial to understanding the dynamics and interplay between the three main dispersal centres that influence the region (northern Pennines, Southern Uplands and Lake District) as well as the Irish Sea Basin and eastern coast of the UK, which were influenced by ice draining away from the central sector of the BIIS. This research question is addressed through mapping of glacial landforms, and investigations of till stratigraphy, sedimentology and provenance.

- ii. *What was the relative chronology of ice sheet advance and decay in the central sector of the BIIS?*

There are currently no dating constraints on the build-up and decay of the central sector of the BIIS. This is fundamental to firstly reconstructing the palaeoglaciology of the ice sheet with respect to ice flow phasing and dynamics, and secondly to assess the relationship of the BIIS to external climate forcings. A chronology of ice sheet build-up and decay is determined by the relative dating of superimposed glacial systems, such as superimposed subglacial lineations, moraines and deglacial depocentres. Furthermore, re-advances and ice-flow phases are correlated with known ice-marginal fluctuations from around the Celtic Sea Basin.

- iii. *Was there evidence for ice stream activity within this sector of the BIIS during the last glacial cycle?*

Ice streams are areas of fast flowing ice within an ice sheet surrounded by much slower moving ice. They drain a disproportionate volume of the ice sheet (90% in Antarctica: Bamber *et al.*, 2000) and thus have a profound effect on ice sheet geometry and stability (Stokes & Clark, 2001). Therefore, the identification of palaeo-ice streams is an important element in reconstructing ice dynamics within the BIIS. Palaeo-ice streams are identified using the criteria outlined by Stokes & Clark (1999).

### 1.3 Region of Interest

#### 1.3.1 Topography, geology and geomorphology

This research is located in north-western England, formerly the central sector of the BIIS, encompassing an area of 11,800 km<sup>2</sup> (Fig. 1.1). The physiography is complex, comprising three major upland regions, the northern Pennines (Alston Block), Lake District and Southern Uplands, separated by the Solway Lowlands, Vale of Eden and the Stainmore and Tyne gaps (Fig. 1.1). The Solway Lowlands form a low lying (< 100 m O.D.) basin opening up westwards into the Irish Sea, while its north, south and eastern ends are bounded by upland terrain (Fig. 1.1). The Vale of Eden trends northwards into this basin, while to the east, the land rises up and over the Tyne Gap (152 m O.D.). The other mountain pass that cuts through the Pennines, the Stainmore Gap, is characterised by a high col (533 m O.D.) (Fig. 1.1).

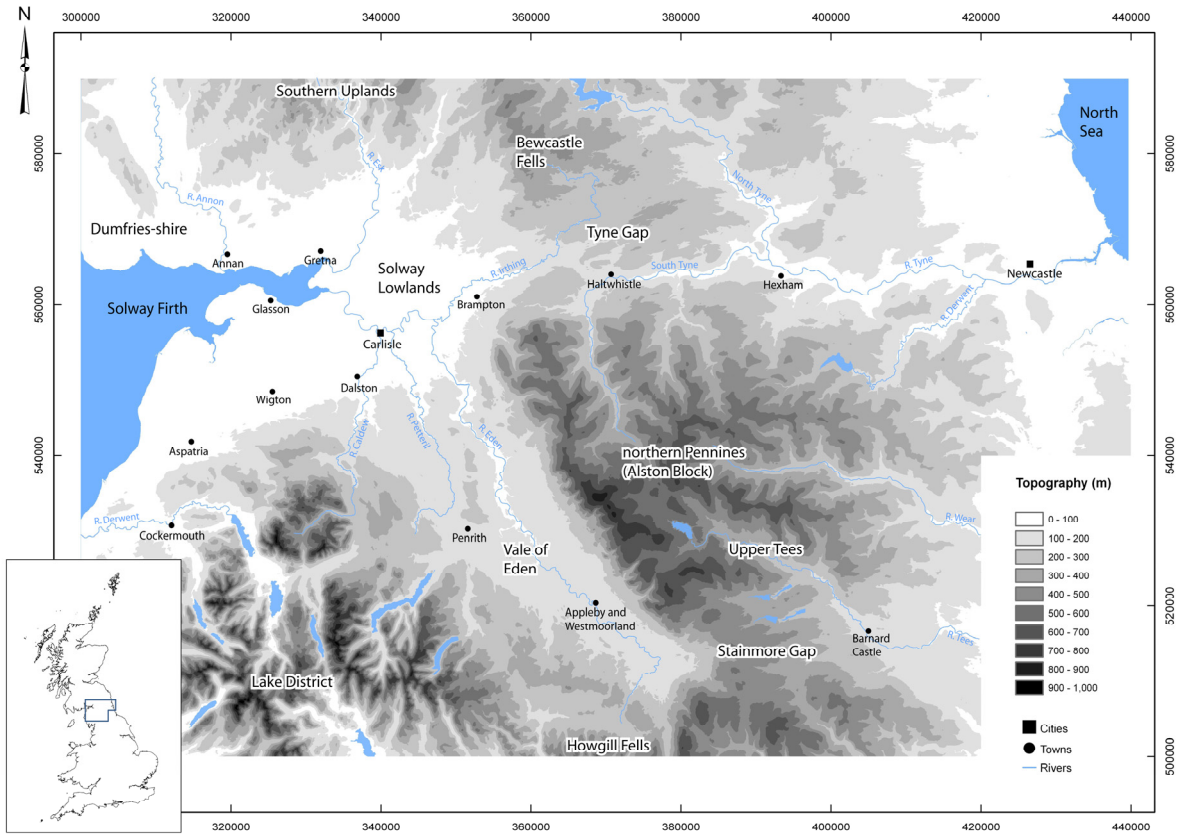
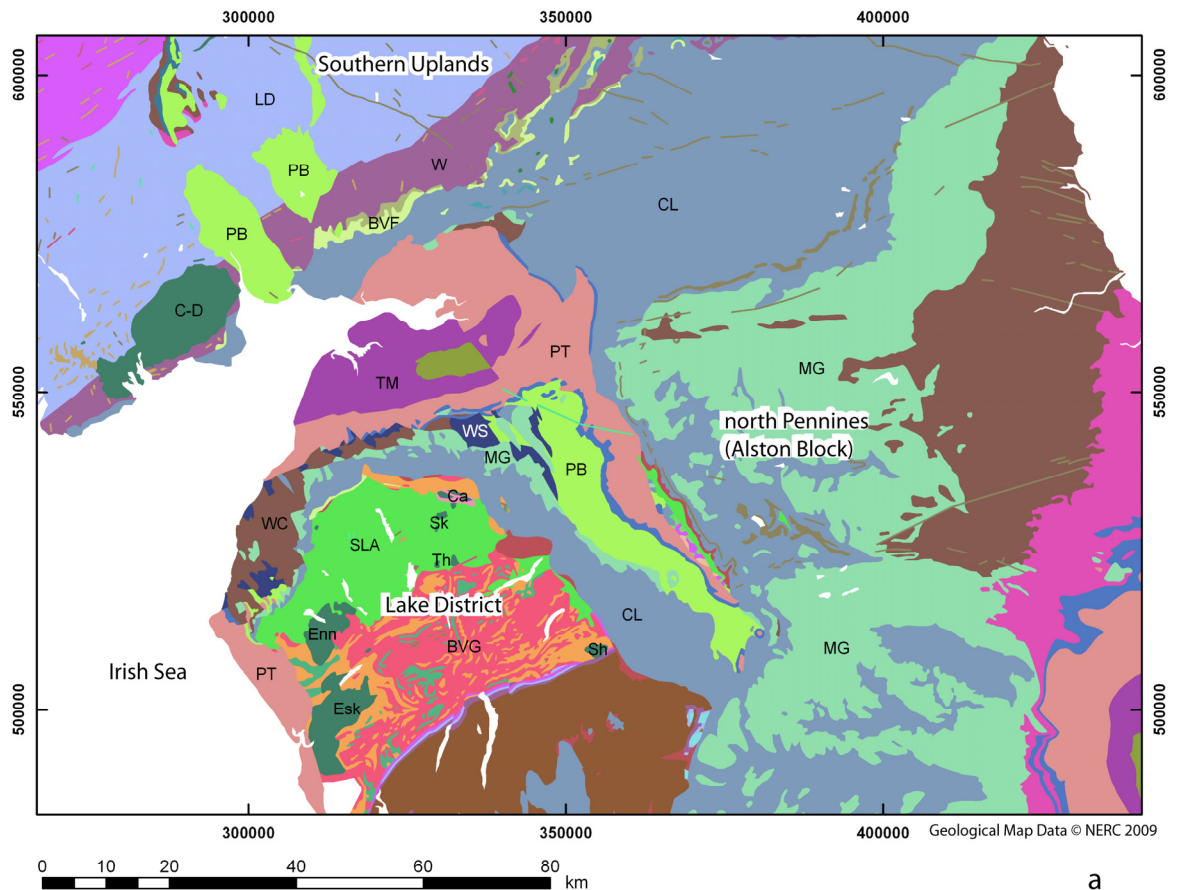


Figure 1.1: Map of the central sector of the BIIS, including topography and places of interest.

This topographic complexity is replicated in the bedrock geology and structure, which has helped define this landscape (Fig. 1.2). The Lake District is formed of a dome of Palaeozoic rocks, comprising from north to south, Skiddaw slates of Ordovician age, the Borrowdale Volcanic series, and Silurian shales, grits, mudstones and siltstones (King, 1979; Fig. 1.2a,b). In parts of the Lake District igneous rocks dominate the landscape, manifest in outcrops of Shap granite, Ennerdale grannophyre, Eskdale granite, Threlkeld syenite and Carrock Fell gabbro (Trotter, 1929; Fig. 1.2a). Likewise, igneous outcrops are also evident in SW Scotland, with the intrusions of Criffel and Dalbeattie granite of particular significance (Charlesworth, 1926; Fig. 1.2a). The Southern Uplands, in SW Scotland is primarily composed of Palaeozoic rocks comprising Ordovician and Silurian grits, slates and greywacke along with infrequent bands of chert. Palaeozoic rocks also form much of the northern Pennines, although they have been subsequently buried by Carboniferous rocks which also occur in the lowland regions (King, 1976; Fig. 1.2a, b). The north Pennine chain dips gently to the east, with its western and northern margins delimited by a fault system which separates it from the Vale of Eden (Pennine fault) and Tyne Gap (Stublick fault) respectively (Fig. 1.2b, c). The Carboniferous rocks of the Tyne Gap (which were deposited in a downwarp called the Northumberland Trough) include rhythmic sequences of limestone, shale, sandstone and coal that dip southwards towards the edge of the northern Pennines (Fig. 1.2). The limestone dies out towards the north, where it is replaced by sandstone and shale (King, 1976).

Bedding structures are particularly apparent in the area from Gilsland eastwards to the river North Tyne, where, along with the Whin Sill, they form prominent north-facing escarpments and southerly dipping slopes (Taylor *et al.*, 1971). A band of the more resistant ‘Whin Sill’ forms a north-facing scarp which runs along the northern-most edge of the Tyne Valley out towards the North Sea (Fig. 1.2c). The fringe of the Vale of Eden and Southern Uplands, along with the south-western edge of the boundary between England and Scotland consists of Carboniferous limestone, which has been subsequently overlain by a triangular wedge of Permo-Triassic ‘new red’ sandstones occupying the Vale of Eden and Solway Lowlands (King, 1979; Fig. 1.2a).



#### Bedrock Geology:

Agglomerate in neck	Llanvirn and Arenig	Rhyolitic and trachytic lava and tuff
Andesitic and basaltic lavas and tuffs	Lower Lias	Rhyolitic lava
Andesitic lava and tuff	Lower Old Red Sandstone	Rhyolitic tuff, including ignimbrite
Ashgill	Lower Westphalian (coal measures)	CL Carboniferous Limestone Series
Ashgill and Caradoc	Ludlow	TM Triassic mudstones
Basal Conglomerate	Ludlow and Wenlock and Llandovery	Tuff
Basalt and spilite	Magnesian Limestone	Tuff and agglomerate (mainly basaltic)
Basalt dolerite and camptonite	Mica-schists	Tuff (mainly andesitic)
Basalt, spilite and rhyolite	Middle Old Red Sandstone	Upper Greensand and Gault
Caradoc	MG Namurian (Millstone Grit Series)	Upper Lias
Diorite and allied intermediate types	PT Permian and Triassic sandstones	Upper Old Red Sandstone
Gabbro and allied types	PB Permian basal breccias	W Wenlock
Granite, syenite and granophyre	Permian mudstones	WC Westphalian Coal Measures
Intermediate and basic rock	Porphyrite, lamprophyre	WS Westphalian and ?Stephanian
Limestone	Rhyolite, trachyte	
LD Llandovery	Rhyolite, trachyte, felsite, elvans	

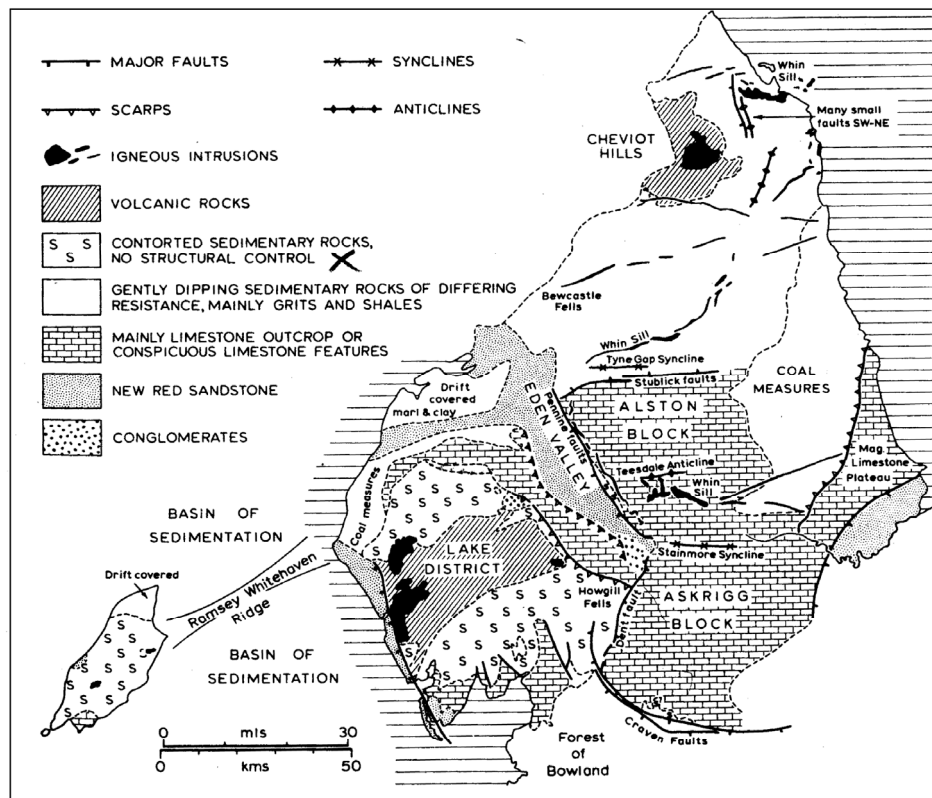
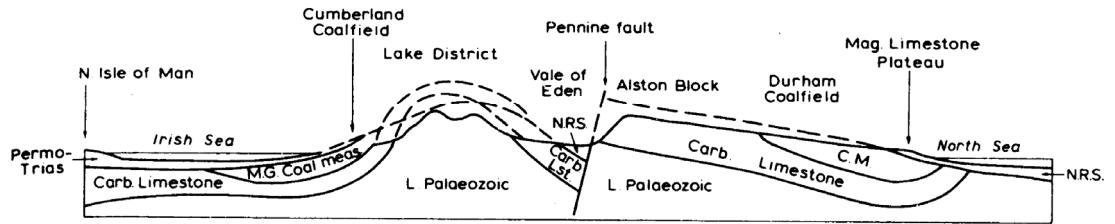


Figure 1.2: (a) 1:625k bedrock geology map of northern England and the Scottish Borders. Granites: Sh = Shap, Esk = Eskdale, Enn = Ennerdale, Th = Threlkeld, Sk = Skiddaw, Ca = Carrock Fell, C-D = Criffel-Dalbeattie pluton. Extrusive igneous: BVG = Borrowdale Volcanic Group, BVF = Birrenwark Volcanic Formation. SLA = Skiddaw Slate Series. Note the Whin Sill orientated SW-NE in the Tyne Gap; (b) generalised geological section running west to east across the region (reproduced from King, 1976); and (c) structural elements in northern England (reproduced from King, 1976).

The geomorphology of NW England and SW Scotland has been heavily influenced by prior glacial activity, with the most recent Late Devensian glaciation accountable for much of the superficial deposits and landforms. The lowland landscape is characterised by a complex and diverse array of

glacial landforms, including subglacial lineations, eskers, glaciofluvial accumulations and meltwater channels (Trotter, 1929, Hollingworth, 1931, Huddart, 1970). Areas of exposed bedrock have been heavily streamlined, while deposition of thick sequences of glacial sediments is commonplace throughout the lowland regions (Huddart & Glasser, 2002).

### 1.3.2 *General palaeoglaciological setting*

Previous reconstructions of the central sector of the BIIS have highlighted the complexity displayed by the glacial landforms and deposits (Trotter, 1929; Hollingworth, 1931). In particular, subglacial lineations indicate multiple ice flow directions often characterised by complete flow reversals (Trotter, 1929; Hollingworth, 1931; Letzer, 1987; Clark R. 2002; Mitchell, 1994, 2007), while meltwater channels and glaciofluvial accumulations demonstrate a complex period of deglaciation characterised by re-advance phases and the extensive build-up of meltwater (Dixon *et al.*, 1926; Trotter, 1929; Hollingworth, 1931; Trotter & Hollingworth, 1932; Huddart, 1970, 1981, 1991; Yorke *et al.*, 2007). Resolving the complexity exhibited within this region of Britain has proved challenging, resulting in a number of glaciologically implausible scenarios and competing theories that fail to fully reconcile all aspects of the glacial geomorphology and sedimentology.

Three distinct phases are currently recognised to have occurred during the last glaciation in the central sector of the BIIS. This includes an early ‘Scottish Advance’, followed by a ‘Main Glaciation’ and then a ‘Scottish re-advance’ phase (cf. Trotter & Hollingworth, 1932). The ‘Scottish Advance’ is delimited by erratic trains, sourced in the Southern Uplands and traced up the Vale of Eden and across the Stainmore Gap. Shap granite erratic trains also indicate the diversion of Lake District ice over the Stainmore Gap during this phase (Trotter, 1929; Hollingworth, 1931; Trotter & Hollingworth, 1932). Erratic trains and drumlins were used to constrain the ‘Main Glaciation’ (e.g. Trotter, 1929). This phase was characterised by northerly moving Lake District ice which coalesced with Scottish ice in the Solway Lowlands before moving eastwards through the Tyne Gap (Trotter, 1929). Westerly moving ice was also recorded in this phase by drumlins trending around the northern margin of the Lake District into the Irish Sea Basin (Trotter, 1929; Hollingworth, 1931). The final phase, which is recorded by a thin till, eskers and delta deposits, marks a temporary re-advance of Scottish ice into the Solway Lowlands during deglaciation (Trotter & Hollingworth, 1932; Huddart, 1970, 1991, 1994).

Deglaciation was characterised by the significant development of ice-dammed lakes within the Solway Lowlands, both during the retreat of the ‘Main Glaciation’ (Huddart, 1970) and the subsequent ‘Scottish re-advance’ (Dixon *et al.*, 1926; Trotter, 1929). These lakes and associated networks of meltwater channels and kame deposits along the Pennine escarpment, in the Brampton region and throughout the Penrith sandstone outcrop have been used to trace the palaeo-margin of the retreating ice sheet (Trotter, 1929; Hollingworth, 1931). Later research re-interpreted many of the meltwater channels as subglacial (Arthurton & Wadge, Huddart, 1970; Greenwood *et al.*,

2007), while the Brampton kame belt was thought to record glaciofluvial deposition on-top of and within stagnant ice (Huddart, 1970, 1981). Meltwater channels that cut across the Tyne Gap watershed, along with a series of glaciofluvial deposits situated in the Tyne Valley demonstrate extensive meltwater drainage (e.g. Trotter, 1929; Yorke *et al.*, 2007), with Glacial Lake Wear dammed up along the eastern margin of this mountainous pass (Teasdale & Hughes, 1999).

## 1.4 Overview of Materials and Methods

### 1.4.1 *Glacial geomorphological mapping:*

#### i. *Conceptual framework*

This work utilises a glacial inversion scheme for the interpretation of landform assemblages mapped from satellite and airborne imagery (Kleman & Borgström, 1996; Kleman *et al.*, 2006; Boulton & Clark, 1990a; Clark, 1997). The theoretical consideration behind this framework is that glacial landforms can be assigned to formative conditions and therefore used to reconstruct the geometry and evolution of former ice sheets. This conceptual framework is now widely applied in the reconstructions of former ice sheets (e.g. Punkari, 1993; Clark *et al.*, 2000; Jansson *et al.*, 2002).

In practice, glacial landforms, such as glacial lineations, eskers, ribbed moraine and meltwater channels are firstly identified and mapped in a GIS package (ArcGIS). At this stage it is important to check against geological structures and human features so that any spurious correlations are removed. The next stage involves the recognition of discrete ‘flowsets’ from glacial lineation patterns. These are grouped according to parallel conformity, morphology and length (Boulton & Clark, 1990a), with each ‘flow set’ interpreted to delimit a distinct ice flow event. An additional grouping scheme also utilised in this thesis, is the identification of ‘fans’, derived on the basis of multiple landform assemblages (Kleman & Borgström, 1996). These can be used to assess internal age chronologies, subglacial thermal regimes, stillstands, ice sheet configurations and the potential locations of ice streams (cf. Kleman & Borgström, 1996). An assessment of whether landforms formed synchronously or time transgressively (internal age chronology) can also be ascertained using ‘flow sets’ (Clark, 1999).

It is now well established that glacial landforms can survive ice sheet overriding, both during dynamic re-adjustments of the ice-sheet system within a single glaciation, and throughout several glacial cycles (Lagerbäck, 1988a,b; Lagerbäck & Robertsson, 1988; Kleman & Borgström, 1990; Clark, 1993; Kleman, 1994). This can result in the cross-cutting and superimposition of glacial landforms (e.g. Rose & Letzer, 1977; Letzer, 1987; Rose, 1987; Salt, 2001; Salt and Evans, 2004; Greenwood & Clark, 2008) which can provide information on the changing ice dynamics through



time (Clark, 1993). These palimpsest glacial geomorphological landforms can also be used to delimit a series of chronologically distinct relative ‘flow phases’ (Boulton & Clark, 1990a, b; Clark, 1993). Cross-cutting patterns in a single glaciation can occur due to ice divide migration, ice stream activation or lobate margin retreat, all of which leave unique geomorphological signatures (Clark, 1997).

ii. *NEXTMap DEM dataset*

NEXTMap DEM was selected because of its high resolution (5 m pixel size). The data comprises a digital elevation model (DEM) derived from Airborne Interferometric Synthetic Aperture Radar (IFSAR). The NEXTMap data has been acquired by the British Geological Survey (BGS) for NERC and is archived at NERC Earth Observation Data Centre (NEODC). Two products of the NEXTMap data were used within this project; a digital surface model (dsm) and a digital terrain model enhanced (dtme) which has been manually edited to remove human features (<http://www.neodc.rl.ac.uk/>). The data is organised into National Grid tiles of 10 x 10km based on Ordnance Survey co-ordinates. Several published cartographic sources and other digital layers were used to supplement and aid the geomorphological mapping. These sources comprise the Glacial Map of Britain (Clark *et al.*, 2004; Evans *et al.*, 2005), bedrock and superficial deposits of Great Britain (DiGMapGB-625 downloaded from the BGS), Ordnance Survey tiles (EDINA digimap) and aerial photographs (Cambridge University Collection of Air Photos (CUCAP), Unit for Landscape Modelling). These layers act as controls on the mapping, provide higher resolution images and are used to identify and eliminate spurious geological or human bias in the mapping.

iii. *Mapped glacial landforms*

Landforms were manually digitised into appropriate vector layers in ArcGIS 9.1. Seven types of landforms were identified and mapped, including: (1) subglacial lineations; (2) hummocky terrain; (3) ribbed moraine; (4) meltwater channels; (5) eskers; (6) glaciofluvial deposits; and (7) transverse ridges. The subglacial lineations were sub-divided into discrete ‘flow-sets’ organised according to its relative age. Furthermore, elongation ratios for the subglacial lineations were calculated. These were displayed as a function of elongation ratio using a colour ramp. Mapping reported here involves approximately 9,000 features.

iv. *Advantages and limitations of glacial geomorphological mapping*

The use of GIS techniques in glacial geomorphology allows for coherent, systematic and efficient mapping over a wide area. It also facilitates the incorporation of several layers of data allowing easy comparisons to be made. The mapping that was carried out utilised standard glacial geomorphological practises and conceptual frameworks (inversion methodology). Finally, being able to map cross-cutting relationships using GIS techniques has provided a rigorous methodology for assessing ice-sheet dynamics at the large-scale.

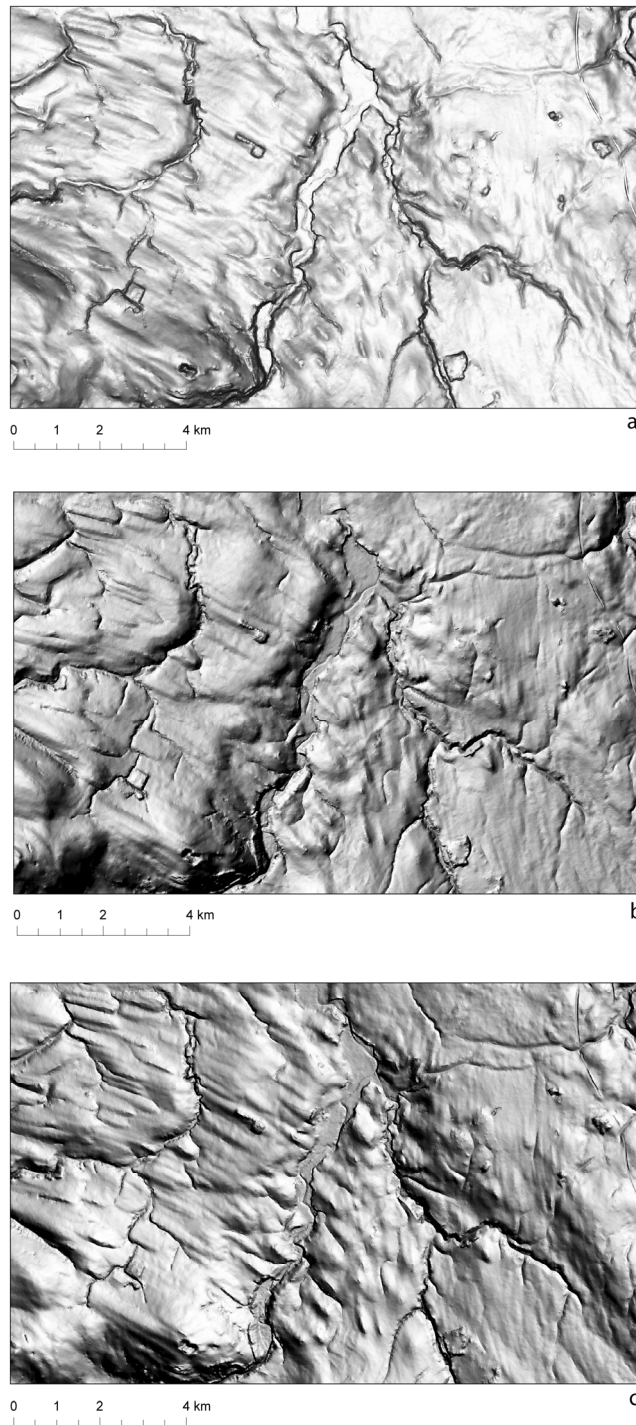


Figure 1.3: Effect of azimuth bias on morphological analysis of glacial lineaments in the Caldew Valley: (a) 'birds eye' view; (b) illumination azimuth: 315°; and (c) illumination azimuth: 045°.

There are several problems inherent within a remote sensing strategy. Smith and co-authors (2001, 2005) emphasize the significance of azimuth biasing, relative size of lineaments and landform signal strength (degree to which the landform can be distinguished from other features by tonal and textural information in the image). The final two are less serious because of the non-selective

nature of bias and the magnitude and well defined character of lineaments (Smith *et al.*, 2001). The orientation of the light source however, can introduce systematic azimuth biasing into the image by either concealing lineaments or making them appear to exist (Smith *et al.*, 2005). The effect of light bias is shown graphically in Figure 1.3. However, this effect has been lessened by repeated mapping from orthogonal light directions (045° and 315°), and also with a ‘bird’s eye’ view to reveal slope curvature (Smith *et al.*, 2005) (Fig. 1.3).

In addition, the identification of ‘fans’ or ‘flowsets’ is essentially a pattern recognition exercise, and as such is open to some subjectivity. It must also be noted that the inversion methodology aims to reduce complexity into a few coherent ‘flow phases’ which are then used to create glacial reconstructions at regional scales. This necessitates some generalisation and simplifying assumptions which can impact upon interpretations at the local scale. It must also be noted that although relative age chronologies can be constructed, absolute ages rely upon corresponding stratigraphic analysis. Coupled to this is the problem of longevity of each interpreted ‘flow phase’, while the co-existence of flow sets is often assigned on the basis of conformity, which relies upon objective reasoning.

#### 1.4.2 *Glacial stratigraphy, sedimentology and provenance:*

##### i. *Glacial stratigraphy*

Exposures of glacial sediments were logged in detail using sketches, vertical lithofacies logs and photographs. Texture, sedimentary structure, colour (Munsell colour chart), bed geometry, contacts, inclusions and deformation structures were all measured and logged, from which lithofacies were identified. Lithofacies codes used in this thesis are based upon those of Evans & Benn (2004). Where similar lithofacies could be ascertained they were grouped into lithofacies associations. Texture was determined in the field, with larger grain-sizes measured and the degree of sorting noted, whilst for smaller-sized grains (sand and below) texture was estimated by rule of thumb (Evans & Benn, 2004). Sedimentary structures including grading and internal bedding (and its inclination) were recorded in the field and depositional structures were also classified according to scale (e.g. ripples, dunes and foresets). Sediments were checked for evidence of fossils, and inclusions detailing evidence of rip-up and mobilization of underlying sediments were noted. The thickness and geometry of beds were measured in detail so information pertaining to the processes and environment of deposition could be ascertained. Contacts between lithofacies, which indicate whether deposition was either continuous or interrupted by episodes of erosion and whether the sedimentary units underwent post-depositional deformation, were also documented. Deformation structures and their timing (syn- or post-depositional) and form (e.g. extensional, pure, sag) provided important additional information on former environments (Benn & Evans, 2004). Scaled

section sketches were drawn at the larger exposures so that the lateral extent of the lithofacies could be assessed. These data were supplemented with borehole logs (BGS) which provided a less detailed but wider coverage, allowing regional stratigraphic correlations.

Stratigraphic logging within the Solway Lowlands was hindered by the lack of suitable exposures. This meant glacial sediments were often observed discontinuously resulting in indirect relationships between individual sections and an incomplete coverage of certain key/complex regions and landforms. Borehole logs allow a better stratigraphic representation of the region although interpretation of this data can be difficult depending on the quality of the original observations made. There is also an inherent bias towards glaciofluvial sediments due to the number of sand and gravel quarries, which provide accessible exposures.

ii. *Clast morphology*

In studies of glacial material clast morphological analysis is a useful method for differentiating between subglacial, englacial and supra-glacial material (Benn, 2004). It can also be used in conjunction with other methods such as clast macro-fabrics, clast lithology and till provenance to work out transport histories and processes at the base of the ice sheet (Benn, 2004). This thesis follows standard procedures outlined by Benn (2004) for both sampling and qualitatively describing the clasts. It is recognised that the techniques involved in clast morphological analysis are subjective and therefore the results are treated with some caution when compared with other studies.

iii. *Macro-fabric data*

Macro-fabric data provides important directional information and can be used to infer depositional processes and deformational mechanisms (e.g. Benn, 1994; Benn & Evans, 1996). Along bedding planes the strike and dip are recorded so as to determine glaciotectionic deformation of Quaternary sediments or depositional surfaces (e.g. delta foresets), while cross-bedding is also recorded in the same way (Benn & Evans, 2004). Measurements of cross-bedding within glaciofluvial sediments can reveal palaeo-flow directions (e.g. ripples) or lateral accretion surfaces (e.g. point bars) (Reineck & Singh, 1975). Pebble imbrication was also used in this study for palaeo-current analysis, with the strike and dip of the a-b plane recorded (cf. Rust, 1972). Faults are recorded by measuring the strike and dip along the fault plane, along with the direction and amount of displacement. Similarly measurements of the strike and dip along the axial plane of fold structures were documented (Evans & Benn, 2004). The orientation of clast fabrics are known to be related to the subglacial stresses that the sediment has undergone and therefore the nature of the deforming bed and strain conditions (Benn, 1994; Hickock *et al.*, 1996; Carr & Rose, 2003; Larsen & Piotrowski, 2003). Sampling and measuring of the a-axis of clasts followed standard procedures

outlined in Evans & Benn (2004). Interpretation of the clast fabric datasets utilised both stereonet plots and eigenvalue analysis (e.g. Benn, 1994; Hicock *et al.*, 1996).

There has recently been some debate as to the usefulness of clast fabrics for distinguishing between different glacial diamicton facies (Bennett, *et al.*, 1999; Evans *et al.*, 2006; Evans *et al.*, 2007). While this argument has merit, as seen by the large range of values and strong overlap between genetic classifications of diamicton, when employed in combination with other stratigraphic and sedimentological criteria it can still provide important information constraining the depositional and deformational processes (Evans & Benn, 2004).

#### iv. *Clast lithology*

Clast lithological analysis is a powerful tool which can be used for stratigraphic correlation and provenance studies (Bridgland, 1986). In particular, the tracing of ‘indicator erratics’ back to their source regions can be used to reconstruct former ice-flow pathways. Sampling procedures are based on those outlined in Bridgland (1986) and Walden (2004), with three size ranges adopted, 8-16 mm, 16-32 mm and 32-64 mm, and a sample size of  $\geq 300$ . Identification of lithologies was carried out using a hand lens and an optical microscope. The main weakness of this technique is the objectivity involved in the identification and classification of the clast lithologies.

#### v. *Thin sections*

Micromorphology is becoming increasingly relevant in glacial sedimentology as it is realised that macro-scale analysis does not always give sufficient information to accurately characterise the depositional or deformational history of many lithofacies (van der Meer, 1993). Micromorphology is the examination of undisturbed sediment using a microscope for structural and compositional evidence of glacial processes through a number of recognisable microstructures (van der Meer, 1987) (Fig. 1.4). Samples, collected in the field using a Kubiena tin, were chosen based on particular structures, lithofacies or boundary contacts. Standard techniques were used for the impregnation and preparation of thin sections as outlined in Murphy (1986) and van der Meer (1993). Analysis was carried out under a petrological microscope using plane and cross-polarised light to view the structural properties of the sediment. Description of the thin sections utilised the approaches outlined in van der Meer (1993), Menzies (2000), Carr (2004) and Hiemstra (2007).

The main source of error in this technique results from sampling or preparation induced microstructures (Carr, 2004). These microstructures, which can result from compaction or deformation during sampling, dewatering during transportation and storage, or impregnation of the resin, must be distinguished from those inherent to the sample itself (Carr, 2004). Additionally, it can prove hard to extract samples if the sediment is fissile or unconsolidated.

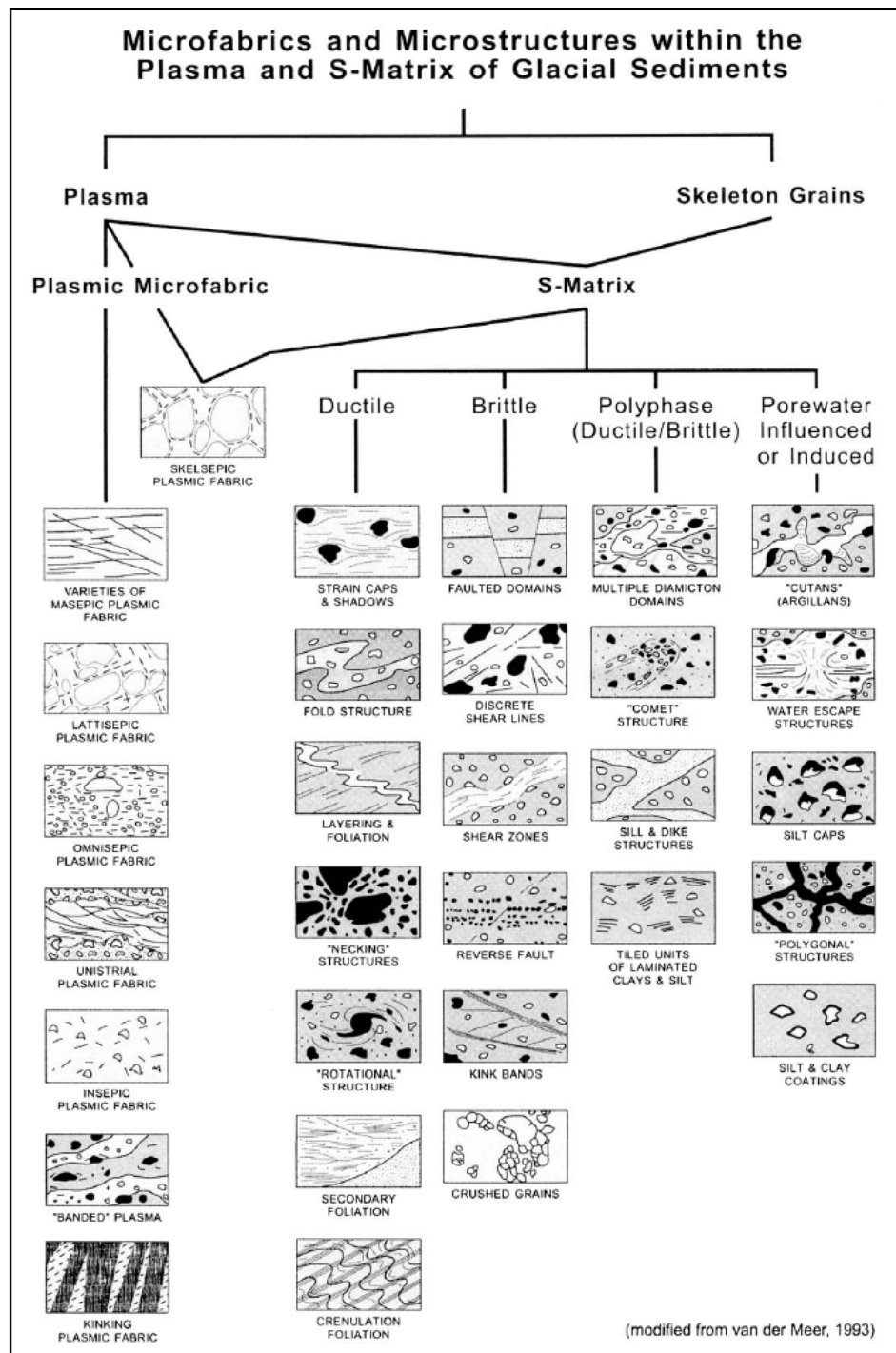


Figure 1.4: Microstructures identified within thin sections of glacial sediments (van der Meer, 1993).

vi. *Particle size analysis*

Particle size analysis is a common technique which can be used to provide information about the source material, as well as the processes of deposition, erosion and transportation (Hoey, 2004). It can therefore be used to distinguish between different flow events and aid in the correlation of lithofacies. Sampling and preparation followed standard procedures outlined by Gale & Hoare,

(1991) and Hoey (2004), with 1 kg of bulk sediment collected at each sampling location. Samples were taken from diamicton lithofacies at 0.5 m intervals, with thinner lithofacies demanding more detailed sampling in order to achieve a representative data set. The  $< 2$  mm fraction, which was separated by sieving (and selected using a riffle box), was analysed using a Beckman Coulter Laser Diffraction Particle Size Analyser, with the results displayed as ternary plots. Two runs were carried out for each sample and an average taken provided that the result graphs were similar. It is recognised that laser coulter typically give results that are systematically different to other techniques (e.g. wet sieving), making cross-study comparisons difficult (Hoey, 2004).

#### vii. *Geochemistry*

Geochemical analysis is a widely used technique for determining the provenance of glacial sediments (e.g. Shilts, 1993). Identification of glacial dispersal trains has allowed ice-flow pathways to be reconstructed (Saarnisto, 1990, Shilts *et al.*, 1979; Dyke *et al.*, 1982; Richards, 2002), while the geochemical signature of sediments can be used to stratigraphically correlate lithofacies (e.g. May & Dreimanis, 1976). Diamictons were sampled at 0.5 m intervals (with smaller exposures necessitating more detailed sampling). The geochemical signature of the glacial sediments was determined using an inductively coupled plasma mass spectrometer (ICP-MS). Details of sample preparation are outlined in Chapter 8, while the errors associated with the ICP-MS are shown in Table 1.1. Multivariate statistical analysis was applied to the geochemical analysis, with both Principle Component Analysis (PCA) and Hierarchical Cluster Analysis (HCA). Both statistical packages are used as an exploratory tool for simplifying the data into meaningful assemblages and hence revealing underlying data structures (Davis, 1986; Temple, *et al.*, 2008). Standardisation and transformation of the datasets were carried out as outlined in Kovach (1995) and Temple *et al.* (2008), while elements with values below the data limit were omitted from the analysis.

The statistical approach to geochemical investigation is an exploratory one in the UK setting and therefore should be used either to formulate hypotheses or as part of a suite of techniques until a denser network of samples is available for the region and clear bedrock sources can be assigned to particular geochemical signatures. These developments will enable the reconstruction of glacial flow patterns to a level compatible with that achieved in North America and Scandinavia.

Element	Average	SD
<i>Li</i>	0.72	0.46
<i>Be</i>	0.17	0.08
<i>B</i>	0.95	0.65
<i>V</i>	0.87	0.69
<i>Cr</i>	2.42	1.18
<i>Mn</i>	15.51	22.75
<i>Co</i>	0.26	0.15
<i>Ni</i>	0.61	0.39
<i>Cu</i>	0.46	0.35
<i>Zn</i>	1.10	0.67
<i>As</i>	0.28	0.52
<i>Se</i>	0.28	0.08
<i>Zr</i>	4.64	1.64
<i>Mo</i>	0.07	0.07
<i>Ag</i>	0.04	0.04
<i>Cd</i>	0.11	0.16
<i>Sb</i>	0.15	0.16
<i>Ga</i>	0.44	0.23
<i>Sn</i>	0.31	0.13
<i>Rb</i>	0.18	0.13
<i>Sr</i>	2.70	1.30
<i>Y</i>	0.05	0.04
<i>Pb</i>	0.47	0.39
<i>Bi</i>	0.06	0.03
<i>Tl</i>	0.05	0.07
<i>Th</i>	0.05	0.07
<i>U</i>	0.05	0.03
<i>Ba</i>	4.93	2.73
<i>Ce</i>	0.15	0.23
<i>Nd</i>	0.18	0.28
<i>Nb</i>	0.26	0.15
<i>Al</i>	333.06	163.47
<i>Ca</i>	132.22	112.41
<i>Fe</i>	471.50	301.96
<i>K</i>	260.09	108.95
<i>Mg</i>	14.67	10.42
<i>Na</i>	98.62	58.68
<i>Si</i>	7429.06	3708.39
<i>Ti</i>	54.34	38.22
<i>P</i>	11.37	4.88

Table 1.1: Laboratory errors ( $\pm$  mg/kg) associated with ICP-MS Total Metal Extraction technique: the reported uncertainty is an expanded uncertainty calculated using a coverage factor of 2 which gives a level of confidence of approximately 95%.



## 1.5 Overview of Results

The results of this work are presented in the following seven Chapters. Chapters 2-3 provide the glacial geomorphological context, while Chapters 4-6 focus on discrete regions and glacial landforms within the study area. Chapter 7 addresses the Scottish re-advance problem, while Chapter 8 attempts to correlate stratigraphic lithofacies, both between fieldsites and down vertical sections, through the novel use of geochemical analysis. Detailed investigations of individual regions and landforms, coupled with regional correlations of flow phases and stratigraphies are combined in Chapter 9 to create a regional synthesis of the glacial history of the central sector of the BIIS.

### 1.5.1 Chapter 2:

Livingstone, S.J., Ó Cofaigh, C. & Evans, D.J.A. 2008. Glacial geomorphology of the central sector of the last British-Irish Ice Sheet. *Journal of Maps*, 2008; 358-377.

This paper describes and presents glacial geomorphological mapping of the area of study. It therefore forms the fundamental framework of the palaeoglaciological reconstruction, with results utilised throughout the remaining chapters.

In this paper I carried out the mapping, writing, figure drawing, image processing and database management. All authors contributed ideas and edited the text. The paper (as published in *Journal of Maps*) has been edited to take into account the most up-to-date mapping, and also to ensure consistency throughout the thesis.

### 1.5.2 Chapter 3:

Evans, D.J.A., Livingstone, S.J., Vieli, A. & Ó Cofaigh C. 2009. The palaeoglaciology of the central sector of the British and Irish Ice Sheet: reconciling glacial geomorphology and preliminary ice sheet modelling. *Quaternary Science Reviews*, 28; 735-757.

This chapter builds on work carried out in the previous article, focusing in on four regions that demonstrate complex, palimpsest flow signatures. These case studies are summarised into four generalised ice flow phases. Forcing of a numerical model of the BIIS facilitates an assessment of the potential for rapid switches in ice flow direction. The implications of this combined geomorphic-modelling approach are: (a) the ice sheet had no real steady state and was therefore subject to constantly migrating ice dispersal centres and ice divides; and (b) subglacial streaming of

flow sets was very dynamic and could be completed by fast flow activity over short time-scales, in some cases less than 300 years.

This paper was compiled by David Evans, with contributions from Stephen Livingstone and Andreas Vieli. The database management, image processing and figure drawing was carried out by Stephen Livingstone (glacial geomorphology) and Andreas Vieli (numerical modelling). All authors contributed ideas and edited the text. The paper (as published in *Quaternary Science Reviews*) has been edited to take into account the most up-to-date mapping, and also to ensure consistency throughout the thesis.

#### 1.5.3 Chapter 4:

Livingstone, S.J., Ó Cofaigh, C. & Evans, D.J.A. in press. A major ice drainage pathway of the last British-Irish Ice Sheet: the Tyne Gap, northern England. *Journal of Quaternary Science*.

This chapter focuses specifically on the Tyne Gap, a mountainous pass through the Pennines, which connects western and eastern England. Three main stages of flow have been demonstrated by glacial geomorphological mapping and detailed sedimentological investigation. Stage I was characterised by convergent Lake District and Scottish ice that flowed east through the Tyne Gap as a topographically controlled ice stream. Evidence of dynamic shifts in ice flow through this stage reflects the change in dominance between Lake District and Scottish ice dispersal centres. Stage II was characterised by the Scottish ice becoming dominant, with ice moving SE down the North Tyne Valley in response to the dissipation of the Solway Firth ice divide and the northwards migration of the ice dispersal centre. Finally stage III saw ice retreat westwards out of the Tyne Gap and the development of a major proglacial drainage network in the Tyne Valley. During the subsequent 'Blackhall Wood re-advance' ice was drawn-down into the Irish Sea Basin. An assessment of the glacial history of the Tyne Gap throughout the Late Devensian has allowed its impact on the palaeoglaciology of the BIIS to be investigated, with its three stages correlated with glacial dynamics both in the Celtic Basin, down the east coast of England and within the Solway Lowlands.

In this paper I carried out the mapping, writing, figure drawing, image processing and database management. All authors contributed ideas and edited the text. This chapter represents an extended version of the paper accepted by the *Journal of Quaternary Science*.

#### 1.5.4 Chapter 5:

Livingstone, S.J., Evans, D.J.A., Ó Cofaigh, C. & Hopkins, J. In press. The Brampton kame belt and Pennine Escarpment meltwater channel system (Cumbria, UK): morphology, sedimentology and formation. *Proceedings of the Geologists Association*.

In this chapter a specific landform-sediment assemblage is investigated, the Brampton kame belt which is situated in the lee of the Pennines in the Solway Lowlands. Glacial geomorphological mapping and sedimentological analysis has allowed a detailed reconstruction of both the morphological features and the temporal evolution of the Brampton kame belt, with processes informed by analogues from modern ice margins. The kame belt demonstrates the development of a complex glacier karst, with glacial meltwater routed both under, through and over dead ice. The system is envisaged to have been dominated by ice-walled lakes and migrating ice-contact drainage networks, with debris flows and the melt out of dead ice becoming major processes during topographic inversion. The Pennine escarpment meltwater channels feed into this kame belt and are composed of a network of anastomosing subglacial channels, and flights of lateral channels. The reconstruction of this landform-sediment assemblage has allowed a detailed conceptual review of the nature of kame genesis, while it has also allowed detailed reconstructions of deglaciation within the Solway Lowlands.

In this paper I carried out the mapping, writing, figure drawing, image processing and database management. Jonathon Hopkins carried out some of the surveying of the Pennine escarpment meltwater channels as part of an undergraduate dissertation. He also contributed Figure 5.7b and supplied the photographs of Figure 5.7c-d. All authors contributed ideas and DJAE and COC helped edit the text. This chapter has been submitted to the journal, *Proceedings of the Geologists Association*.

#### 1.5.5 Chapter 6:

Livingstone, S.J., Ó Cofaigh, C., Evans, D.J.A. & Palmer, A. In press. Sedimentary evidence for a major glacial oscillation and pro-glacial lake formation in the Solway Lowlands (Cumbria, UK) during Late Devensian deglaciation. *Boreas*.

In this paper sedimentological investigation of Late Devensian glacial deposits from the Solway Lowlands are used to reconstruct ice sheet dynamics. Specifically, laminated sediments sandwiched between diamicton lithofacies are interpreted as glaciolacustrine varves. These indicate the former presence of a proglacial lake which existed for  $\geq 261$  years and covered an area of at least 140 km<sup>2</sup>. Further glaciofluvial sediments found throughout the Solway Lowlands demonstrate the formation of an ice-free enclave, constrained by ice within the Irish Sea Basin. Re-advance of ice glaciotectionised the sediments, deposited a capping till and drumlinized the lowland region north of the Lake District. This paper therefore reports the discovery of a hitherto unrecorded re-

advance within the Solway Lowlands. Geomorphological evidence indicates that this re-advance occurred prior to the Scottish re-advance, while detailed cross-basin correlations allow it to be tentatively assigned to the Gosforth oscillation.

In this paper I carried out the mapping, writing, figure drawing, image processing and database management. Adrian Palmer produced the thin sections of the laminated sediments. All authors contributed ideas and edited the text. This chapter has been submitted to the journal *Boreas*.

#### 1.5.6 *Chapter 7:*

Re-advance of Scottish ice into the Solway Lowlands (Cumbria, UK) during the Main Late Devensian deglaciation.

This chapter combines geomorphological and sedimentological investigations to critically review evidence pertaining to the re-advance of Scottish ice into the Solway Lowlands. A large ice-contact delta at Holme St Cuthbert, eskers along the Lias Plateau and sedimentological evidence of a thin upper till overlying glaciofluvial and glaciolacustrine deposits are all attributed to the Scottish re-advance. The Holme St Cuthbert delta marks a major still-stand position, which led to the formation of a large pro-glacial lake in the region of Wigton, dammed up against ice impinging onto the Cumbrian coast. Evidence for an extensive re-advance into the Solway Lowlands is now thought to be incongruous, with much of this complex glacial stratigraphic and geomorphological evidence reconciled with the earlier 'Blackhall Wood' re-advance. However, the subdued eskers and thin discontinuous upper till sheet does support the notion of a thin, transient advance into the Solway Lowlands. A Lake District re-advance signal is identified in the northern sector of the Vale of Eden, although it is not known whether this formed part of a synchronous pan-Cumbrian re-advance or a separate internal re-configuration.

In this paper I carried out the writing, mapping, figure drawing, image processing and database management. DJAE and COC helped edit the text.

#### 1.5.7 *Chapter 8:*

The application of geochemistry and particle size analysis to the investigation of glacial stratigraphy in the central sector of the last British-Irish Ice Sheet.

The use of geochemical investigations to constrain ice-flow histories has been successfully demonstrated in Fennoscandia and North America, yet only sporadically utilised within the U.K. This chapter applies geochemical and particle size analysis, and exploratory statistical

techniques to the investigation of glacial stratigraphy in the glacially and geologically complex Solway Lowlands and Tyne Gap region of NW England. The techniques used in this chapter successfully distinguish and, therefore, verify stratigraphic units identified independently using geomorphological and sedimentological methods. Five major geochemical assemblages are recognised from this analysis, which discriminate between Scottish, Lake District, mixed Lake District and Scottish and local provenances. These assemblages are used to constrain ice-flow histories within the central sector of the BIIS.

In this paper I carried out the mapping, writing, figure drawing, image processing and database management. DJAE and COC helped edit the text.

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## Chapter 2

# Glacial geomorphology of the central sector of the last British-Irish Ice Sheet

Livingstone, S.J., Ó Cofaigh, C. & Evans, D.J.A. 2008. Glacial geomorphology of the central sector of the last British-Irish Ice Sheet. *Journal of Maps*, 2008; 358-377.

### Abstract

We here present a glacial geomorphological map (see back of thesis) covering 11,800 km<sup>2</sup>, at a scale of 1:550,000, of the central sector of the last (Main Late Devensian) British-Irish Ice Sheet. The map is based on the 5 m resolution NEXTMap dataset. Seven landform types have been mapped; subglacial lineations, hummocky terrain, transverse ridges, ribbed moraine, meltwater channels, eskers and glaciofluvial sediment accumulations. The subglacial lineations have been further sub-divided into a series of 'flow sets' based on their morphology, conformity and length. Over 9,000 individual landforms have been identified within the study area, concentrated predominantly in the lowlands of the Vale of Eden, Solway and over the Tyne and Stainmore Gaps. A palimpsest geomorphic signature characterised by cross-cutting flow-sets is interpreted as evidence that dynamic, multiphase ice-flow occurred throughout the Main Late Devensian (Marine Isotope Stage 2) in response to migrating ice dispersal centres and ice divides. A relative chronology has been constructed and interpreted, based on the complex, cross-cutting flow signatures displayed throughout the region.

## 2.1 Introduction

The area formerly covered by the central sector of the last British-Irish Ice Sheet (BIIS) encompasses the Solway Lowlands, Vale of Eden, Stainmore and Tyne Gap regions (Fig. 2.1) and contains a complex palimpsest of subglacial bedforms formed during the Main Late Devensian (Marine Isotope Stage 2) glaciation. Previous glacial geomorphological mapping of the region (e.g. Trotter, 1929; Hollingworth, 1931; Letzer, 1978; Smith, 2002) highlighted the complexity of landforms relating to competing ice dispersal centres in the Southern Uplands, Lake District, Irish Sea and Pennines. However, the compilation of data for the Glacial Map of Britain (Clark *et al.* 2004; Evans *et al.* 2005) revealed incomplete coverage and a lack of knowledge concerning the relative and absolute timing of Late Devensian glacial events in the central sector of the BIIS. This is important because the ice-flow from the central sector of the BIIS influenced and/or controlled

flow in key areas of the ice sheet such as the Irish Sea Basin (e.g. Roberts *et al.* 2007), the Southern Uplands (e.g. Salt, 2001) and the English east coast (e.g. Raistrick, 1931; Davies *et al.*, 2009).

Palaeoglaciologists now routinely acknowledge that ice-sheet flow during a glacial cycle can be highly dynamic (e.g. Dyke & Morris, 1988; Boulton & Clark 1990a, b; Clark, 1997, Clark & Meehan, 2001) with several phases of flow often being preserved as a series of cross-cutting and superimposed landforms. This paper presents the glacial landform evidence of the central sector of the BIIS during the Main Late Devensian focusing on the location of palaeo-ice divides, dispersal centres, flow trajectories and relative flow phasing through the mapping of subglacial bedforms and identification of cross-cutting relationships.

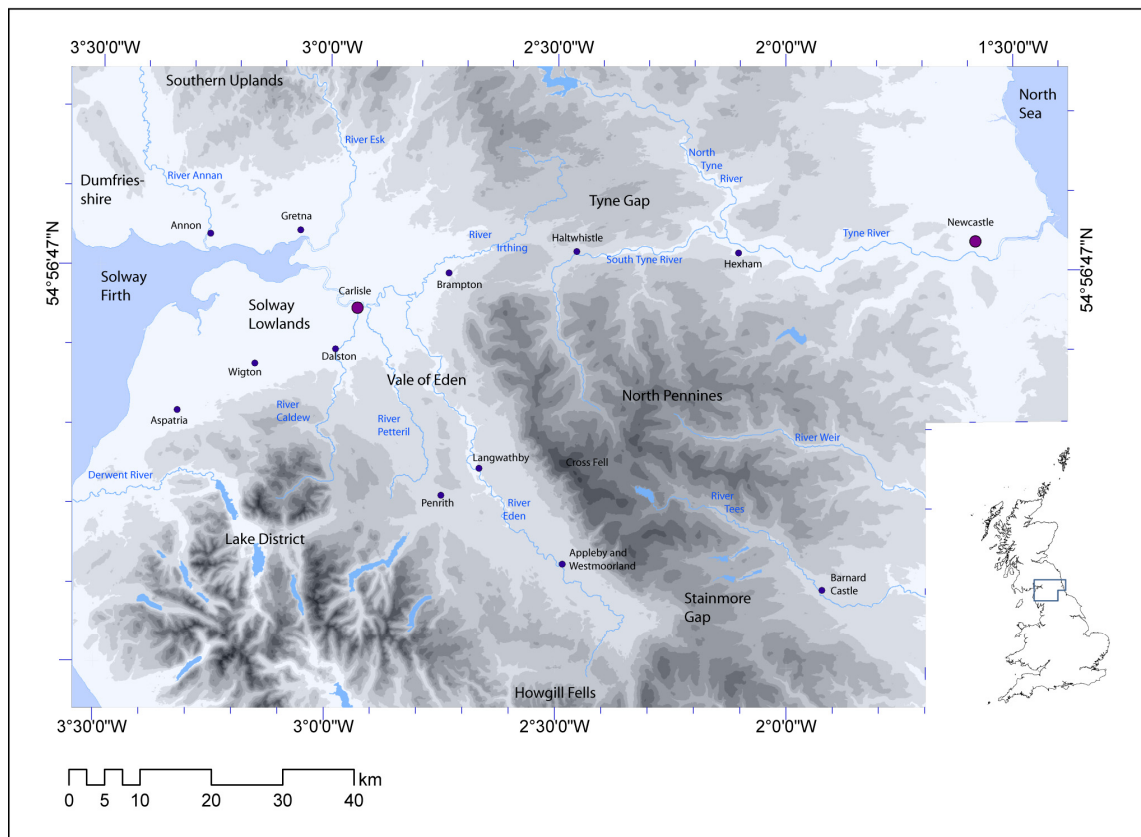


Figure 2.1: Location and topography of the study area.

## 2.2 Methods

### 2.2.1 Dataset:

The glacial geomorphology was mapped using NEXTMap data, which covers all of England, Wales and Scotland. These data comprise a digital elevation model derived from airborne Interferometric Synthetic Aperture Radar (IFSAR), which has enabled rapid assessment of the geomorphology at high spatial resolutions. The data are organised into National Grid tiles of 10 x 10 km based on Ordnance Survey co-ordinates. These data have been acquired by the British



Geological Survey (BGS) for NERC and are archived at NERC Earth Observation Data Centre (NEODC). Two products of the NEXTMap data were used within this project; a digital surface model (DSM) and a digital terrain model enhanced (DTME) which has been manually edited to removed human features. Both datasets have a spatial resolution of 5 m (<http://www.neodc.rl.ac.uk/>). In addition, maps of the bedrock geology and superficial deposits (DiGMapGB-625 downloaded from the BGS) were overlain onto the DSM and DTME.

Smith & Clark (2005) and Smith & Wise (2007) identified a series of problems with mapping glacial landforms from remotely sensed datasets including azimuth biasing, relative size of bedforms and landform signal strength (the degree to which individual landforms can be distinguished from other features). It has been estimated by Smith *et al.* (2006) that NEXTMap provides the most complete dataset for mapping, but that, in comparison to field mapping only visualises about 50% of what is on the ground. The effect of light biasing has been lessened by repeated mapping from two orthogonal light directions and a ‘bird’s eye’ view which reveals slope curvature by shading and flat areas as light (Smith & Clark, 2005).

#### 2.2.2 Mapped landforms:

The Table below lists the glacial landforms which have been identified and mapped. Research theory, methodology and a descriptive overview of the landforms within the study area are all discussed in this section.

Landform type	Number mapped
Subglacial lineations	8050
Hummocky terrain	102
Ribbed moraine	18
Meltwater channels	800
Eskers	13
Glaciofluvial sediment accumulations	36
Transverse ridges	3

Table 2.1: Mapped glacial landforms

##### i. *Subglacial lineations:*

Subglacial lineations are used to reconstruct the glaciodynamics of the BIIS with respect to ice flow orientations, directions and relative velocities (Clark, 1997). 8050 lineations have been mapped within the field area with coverage ubiquitous throughout the lowlands (predominantly below the 400 m contour). The density of coverage varies, with sparse groups both in the deglacially dominated terrain of the Solway Lowlands, and towards the eastern end of the Tyne Gap where they become more subdued. The most densely populated regions are within the Tyne Gap and Vale of Eden. An example of these data is presented in Figure 2.2A which shows a

heavily lineated area in the Vale of Eden with a southeast to north-westerly orientation of lineations. Streamlined landforms include drumlins, flutes, crag-and-tails and lineations which are interpreted as forming subglacially, parallel to ice flow. These features can be either constructional (till deposition) or erosional (bedrock moulded, or till eroded). It is difficult to determine, based on the geomorphology alone, which category particular lineations belong too. However, when combined with the mapped superficial deposits (BGS DiGMapGB-625), areas of bedrock moulded lineations have been picked out, such as in the Tyne Gap. Care has been taken when interpreting these bedrock moulded lineation as geological structure can influence landforms orientation and therefore its 'flowset' grouping. In order to avoid this problem, lineations which are clearly affected by bedrock structure (from overlays of superficial and geological maps) have not been used in the reconstruction of flow sets and flow phasing (see Chapters 3 & 4 for more detailed comments). The lineations have been grouped into discrete 'flow sets' (Table 2.2), defined as a collection of glacial features formed during the same flow phase and under the same conditions (Clark, 1997, 1999). The flow sets are defined on the basis of conformity, length, parallelism and morphology (Clark, 1999). Figure 2.2B displays a series of cross-cutting 'flow sets' west of the Stainmore Gap. The direction of flow is identified by observation of the stoss (up-ice) and lee (down-ice) forms of drumlins. In this case (Fig. 2.2A) the steeper stoss ends to the southeast and tapered lee ends to the northwest indicate former ice flow was to the northwest. Elongation of the lineations has been quantified (length/width) thereby allowing an approximation of relative velocity which can be assessed both spatially and temporally (Stokes & Clark, 2001; Stokes & Clark, 2002).

ii. *Hummocky terrain:*

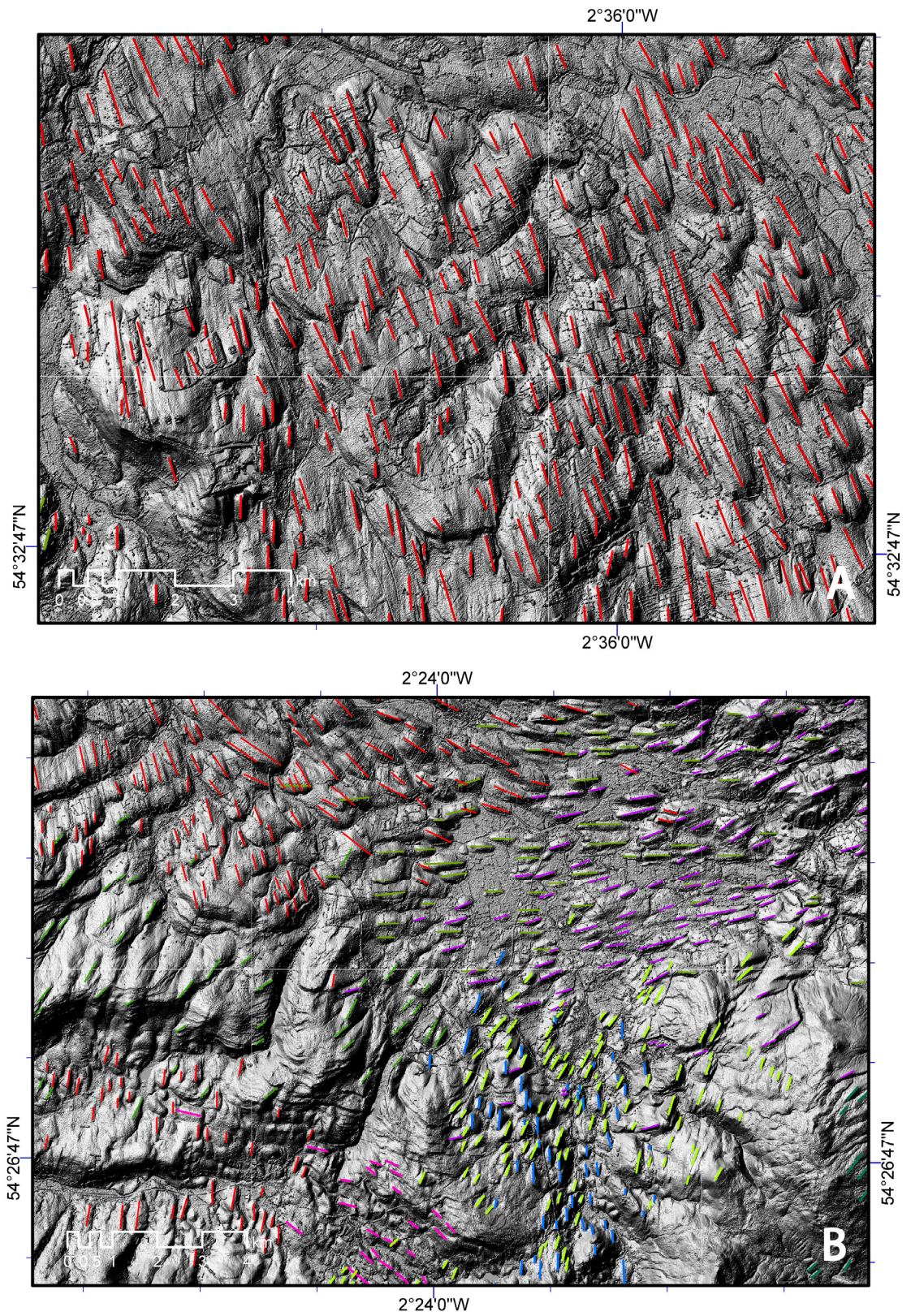
Subglacially formed hummocky terrain is identified as irregular morainic topography characterised by elongation ratios of *ca.* 1:1 and therefore, indistinct orientations (as illustrated in the Solway Lowlands; Fig. 2.2C). Hummocky terrain is commonly associated with stagnant down-wasting ice (e.g. Eyles *et al.*, 1999; Boone & Eyles, 2001). The 102 mapped hummocky terrain landforms are situated within the Solway Lowlands, formed in close association with ribbed moraine, at the northern margin of a flow set arcing out into the Irish Sea Basin (see map, Fig. 2.2C). The sequence of subglacial hummocky terrain and ribbed moraine thus seems to depict the lateral margin of the flow set, where inferred ice flow velocities were significantly reduced.

iii. *Ribbed moraine:*

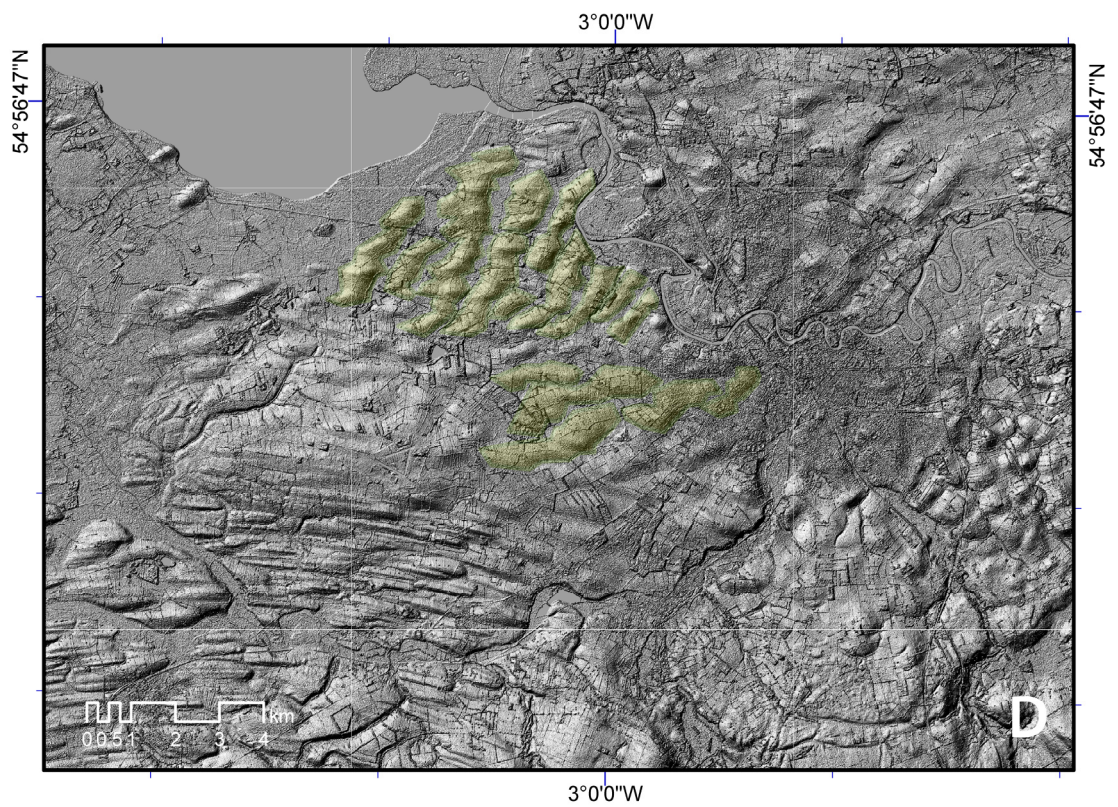
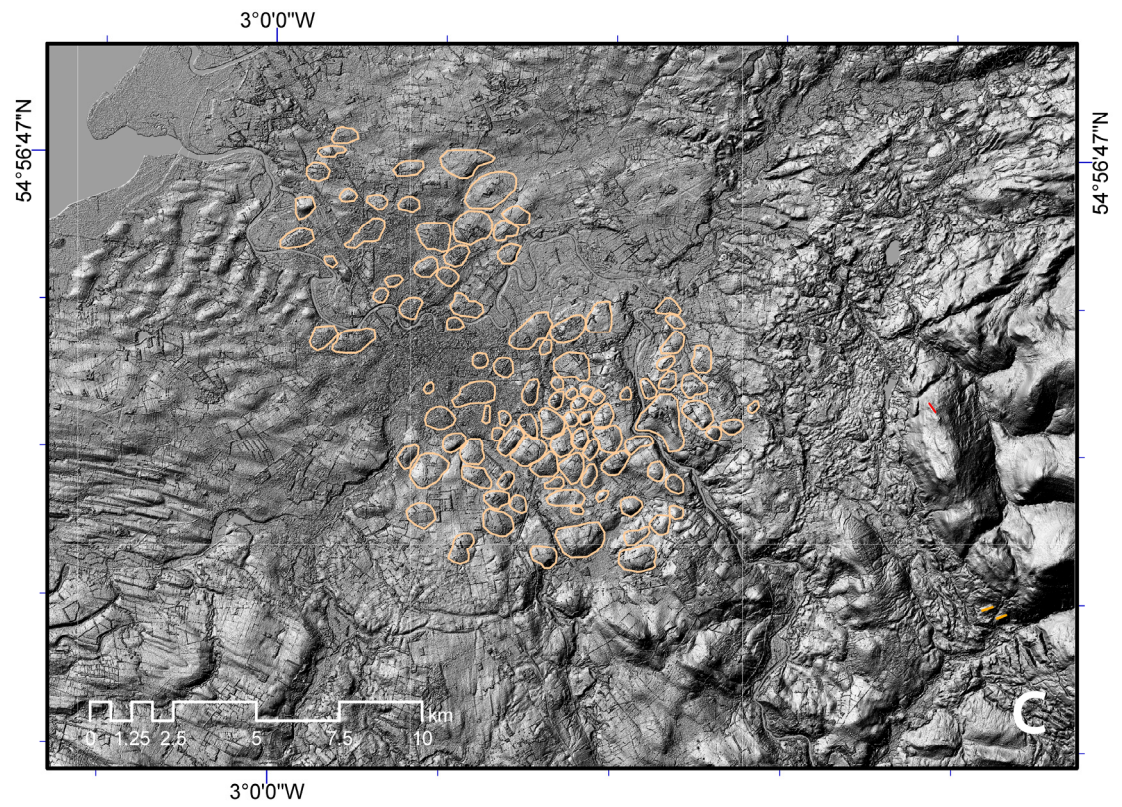
Ribbed moraines are subglacial ridges formed transverse to ice flow. They display a wide range of shapes and patterns and often have drumlinised ridges and transitional associations with lineations (Dunlop & Clark, 2006). This is evident in Figure 2.2D in the Solway Lowlands, east of Carlisle, where drumlins partially overprint the ribbed moraine and a northwards transition occurs between elongate lineations to the south, which grade into more subdued and less elongate lineations and



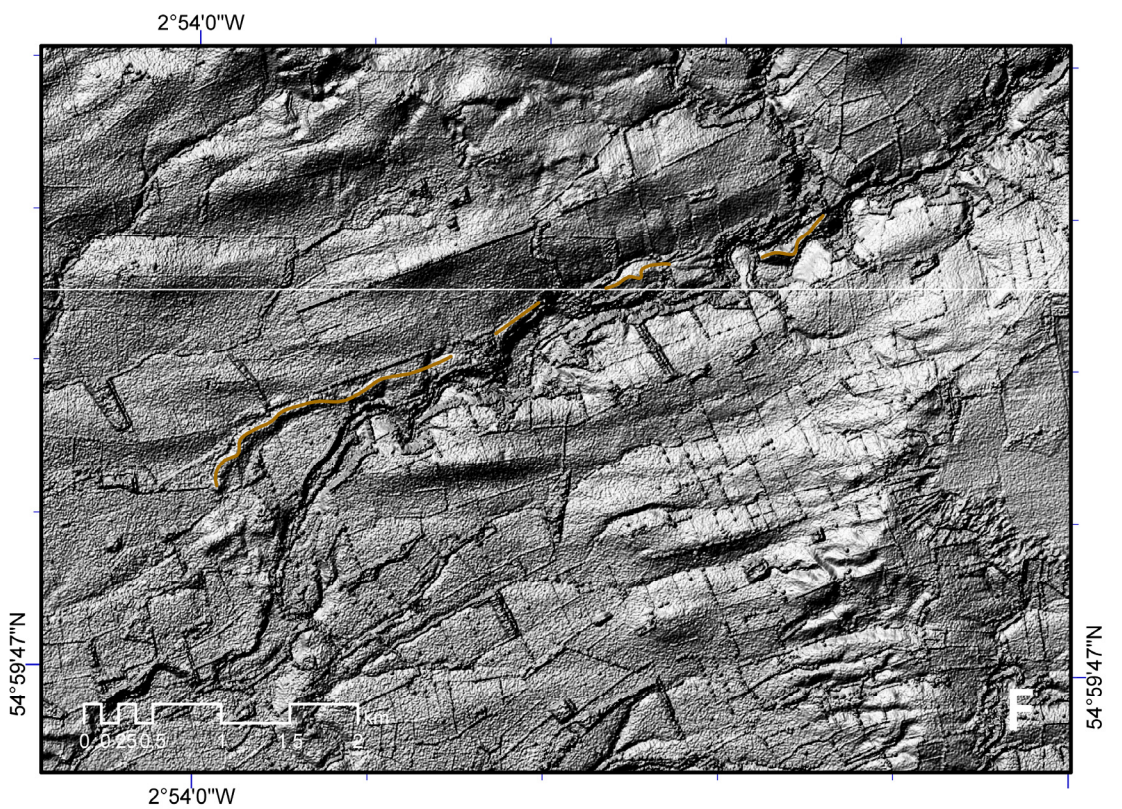
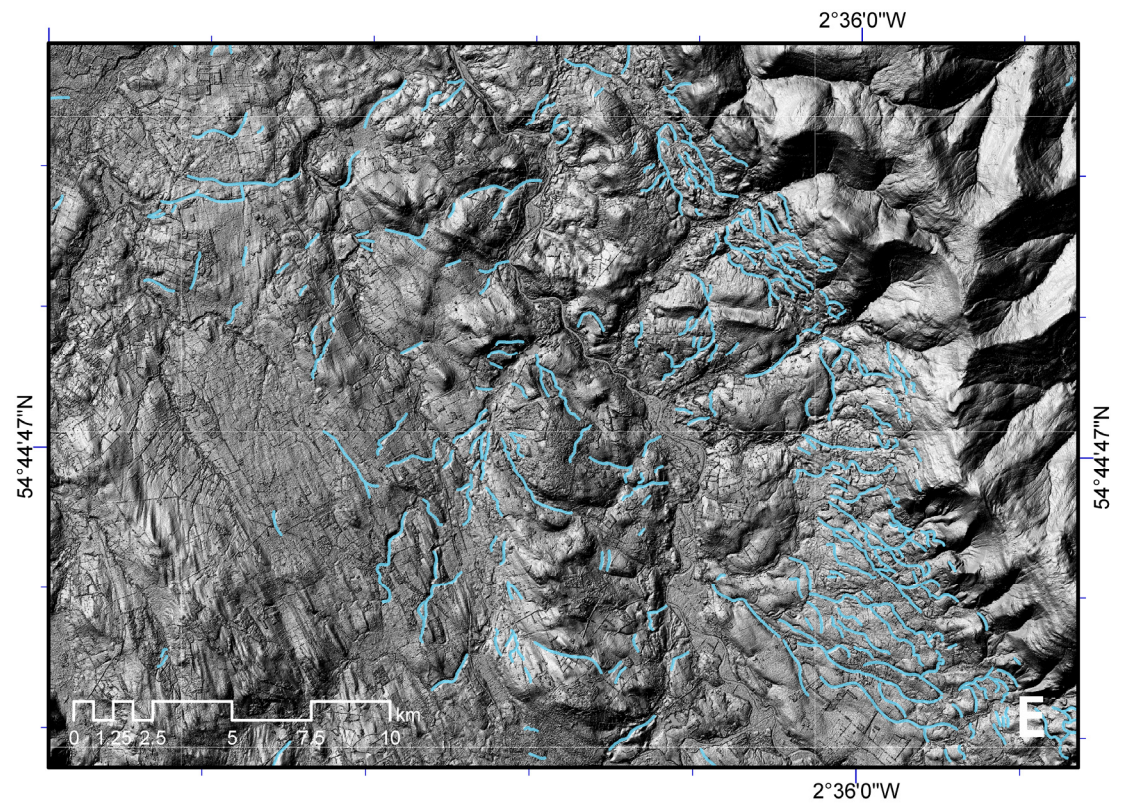
then into ribbed moraine. This transition is evident further eastwards, where the drumlins grade into hummocky terrain (Fig. 2.2C).













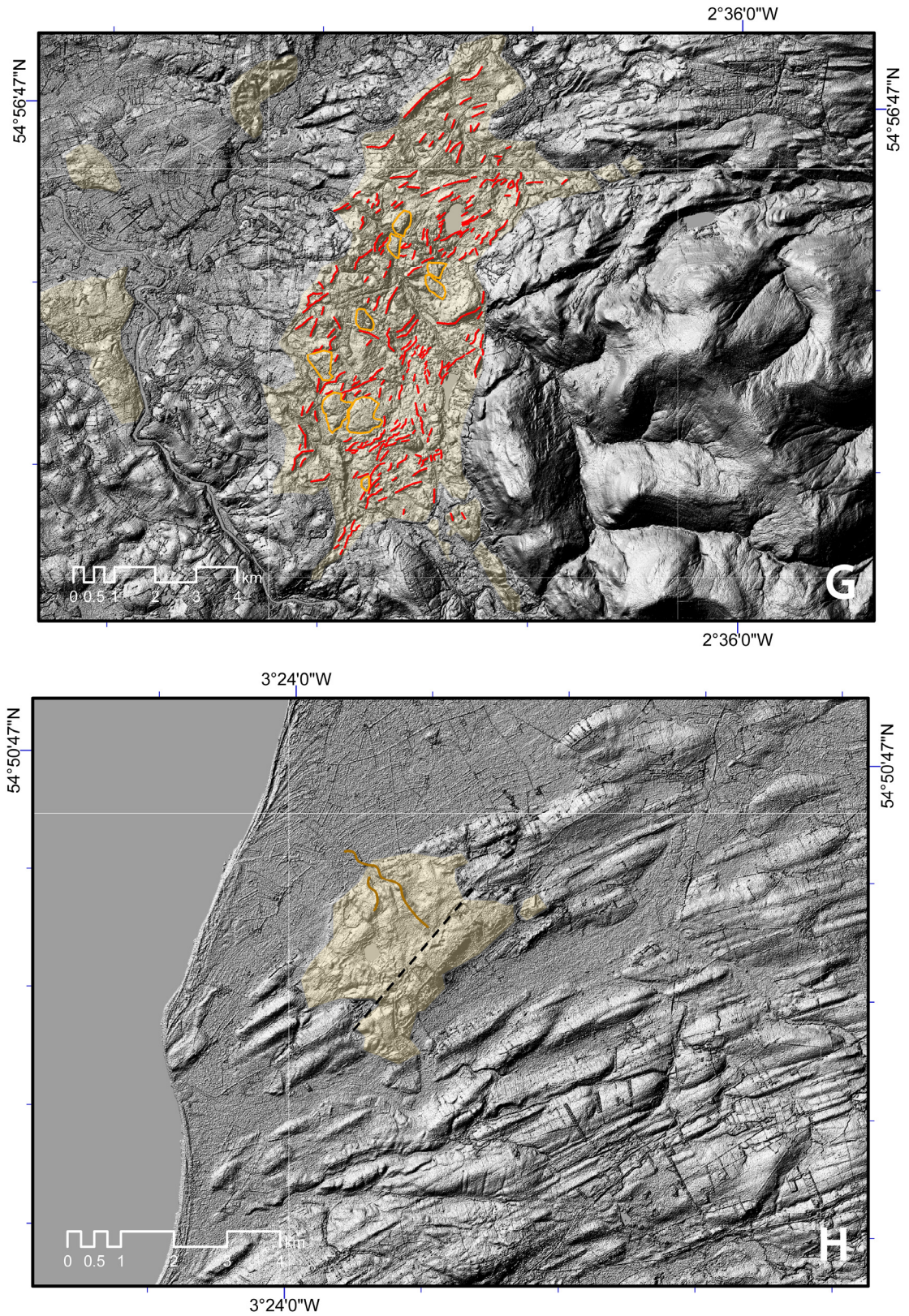


Figure 2.2: Examples of glacial landforms: 2A, SE-NW orientated drumlins in the Vale of Eden; 2B, a series of cross-cutting 'flow sets' to the southwest of the Stainmore Gap. Flow sets are distinguished based on the criteria of Clark (1999). 2C, hummocky moraine in the Solway Lowlands; 2D, Ribbed moraine with drumlinised ridges to the west of Carlisle; 2E, SE-NW parallel-orientated meltwater channels on the Pennine escarpment, probably formed at the ice margin due to progressive ice-sheet surface-lowering and a series of

< Figure 2.2 (continued)... channels crossing the Penrith sandstone ridge related to a series of eskers and delta formations; 2F, sinuous, beaded esker in the Lyne Valley orientated in a SW-NE direction; 2G, Brampton kame belt 'hugging' the northwest corner of the Pennines and containing a series of SW-NE orientated ridges (red), flat topped hills (green polygons) and kettle holes; 2H, ice-contact delta at Holme St Cuthbert. The ice contact face is demarcated by the dotted black line, with eskers and dead ice features to the northwest indicating the ice marginal zone. The delta would have been fed by ice flow from the northwest. Relief shaded DEM (azimuth: 315).

iv. *Meltwater channels:*

Meltwater channels are an important record of the pattern of ice sheet retreat. However, it is often difficult to distinguish between features produced predominantly by post-glacial fluvial activity, which now dominates, and the relict meltwater channels formed during deglaciation. Holocene activity has to a certain extent been influenced by relict channels with modern streams re-occupying previous meltwater channels. Furthermore, being able to elucidate whether meltwater channels formed subglacially, proglacially or ice marginally is crucial in distinguishing different patterns of ice sheet retreat. Therefore, the interpretation of meltwater channels relies on identifying whether they formed sub-aerially, sub-marginally or sub-glacially, and also differentiating between fluvial and glacial channels. The criteria of Greenwood *et al.*, (2007) are used for this. 800 meltwater channels have been mapped in the study area. Especially dense channel networks are mapped along the Pennine escarpment (Fig. 2.2E), the Penrith sandstone outcrop (Fig. 2.2E), the ridge running south-west of Appleby-Westmorland (see map), the northern slopes of the Pennines (see map) and to the north of the Lake District (see map). A series of distinctive drainage network morphologies are also displayed, ranging from dendritic systems (southern ridge of the Vale of Eden) to parallel channels (northern slopes of Alston Block), anastomosing arrangements (lower Tees valley), major spillways (Dalston Gap) and isolated channels (throughout Edenside and the Tyne Gap). Channels intimately linked to lineation patterns have also been noted in areas such as the North Tyne Valley while in other areas some channels are situated transverse to lineations (Solway Lowlands). Thus, the central sector of the BIIS was associated with a myriad of meltwater channels which have been classified and used to demarcate ice-marginal positions during deglaciation.

v. *Eskers:*

Eskers are elongate, sinuous ridges of glaciofluvial sand and gravel (e.g. Fig. 2.2F in the Lyne Valley), that generally form parallel to the direction of ice flow (Flint, 1971; Benn & Evans, 1998). This means they are normally concurrent with the lineations with which they are juxtaposed, although divergence can occur locally due to a time separation between the formation of the lineations and eskers (Kleman & Borgström, 1996). Eskers form time-transgressively, close to a

retreating margin (Kleman & Borgström, 1996) either supraglacially, englacially or subglacially (Price, 1973). Well developed networks of eskers are normally restricted to wet-bedded areas of glaciers (Kleman & Borgström, 1996). Of the 13 eskers that have been mapped within the study area, 4 are intimately associated with glaciofluvial sediment accumulations, for example at the Holme St Cuthbert deltaic complex (Fig. 2.2H) and the Brampton kame belt (Fig. 2.2G) (Huddart & Glasser, 2002). Other examples of eskers within the field area demarcate complex patterns of lobate retreat and stagnation, such as around the Penrith sandstone ridge (Trotter, 1929; Huddart & Glasser, 2002).

vi. *Glaciofluvial sediment accumulations:*

Glaciofluvial sediment accumulations cover a wide range of discrete landforms including, sandur deposits, kames, relict deltas and outwash plains. 36 accumulations of sand and gravel have been mapped within the study area (based on both the morphology and the BGS map of superficial deposits). Ice-contact deltas such as at Holme St Cuthbert and Baronwood (Huddart & Glasser, 2002) are characterised by a steep ice contact face and internally have distinctive sedimentology and stratigraphy (Fig. 2.2H; Holme St Cuthbert). The Brampton kame belt displays very irregular topography containing kettle holes, ridges and flat-topped hills (Fig. 2.2G), indicating that formation occurred during a complex period of stagnant, *in situ* downwasting (Huddart & Glasser, 2002).

vii. *Transverse ridges:*

Subglacial lineations are occasionally associated with discrete transverse ridges, formed near to, or at the margin of a 'flow set'. This landform-type is limited to the Tyne Gap, with three NNW-SSE orientated ridges identified within 18 km of the east coast. These landforms are identified as moraines, formed at still-stands during deglaciation and thought to constrain the time-transgressive formation of the WSW-ENE orientated 'flow set' (cf. Clark, 1999), along with a series of meltwater channels and kame deposits (Smythe, 1912).

### 2.2.3 *Identifying complex flow:*

There is a wide body of literature which recognises crossing or superimposed lineations that have been preserved beneath ice sheets during the Quaternary (Boulton & Clark, 1990; Letzer, 1987; Rose, 1987; Rose & Letzer, 1977; Salt, 2001; Salt & Evans, 2004). These superimposed subglacial landforms survive because of the ability of ice sheets to preserve geomorphological flow features (Kleman, 1994). The co-existence of landforms superimposed at different orientations represents a palimpsest of different ice flow events and, therefore, a series of chronologically distinct relative flow phases (Boulton & Clark, 1990a, b; Clark, 1993). In terms of ice-flow evolution, cross-cutting



lineations can be interpreted in terms of migrating ice divides, ice stream activation and lobate retreat (Clark, 1997) all of which leave unique geomorphological signatures. Cross-cutting may also relate to formation during multiple glaciations although this is unlikely to be the case in the present study as the entire field area was glaciated during the Late Devensian (Evans *et al.*, 2005). This is supported by the regional chronostratigraphy, with well dated events such as the Dimlington stadial (Bateman *et al.*, 2008) and extension of the Irish Sea Ice Stream to its maximum extent (e.g. Ó Cofaigh & Evans, 2007) able to provide constraints on key flow phases (developed further in Chapter 9). This is best exemplified by the flow of ice through the Tyne Gap (see Chapter 4), which has been correlated to the Blackhall till at Whitburn Bay and assigned to the Dimlington stadial (Davies *et al.*, 2009). Similarly flow of ice through the Stainmore Gap can be correlated with formation of the Escrick moraine in the Vale of York, which together with the east coast ice, resulted in the formation of Lake Humber, which has been dated to the Late Devensian (Catt, 2007). In addition, a peat layer found beneath till at Scandal Beck at the head of the Eden Valley is dated to the Ipswichian interstadial (Carter *et al.*, 1978), while the majority of glacial deposits have been assigned to the waxing and waning of a single ice sheet (Huddart & Glasser, 2002; Stone *et al.*, in press). By unravelling the complex palimpsest signature, through cross-cutting relationships identified between ‘flow sets’, a model of dynamic, multiphase ice flow can be reconstructed throughout a single glacial cycle.

Complex flow has been identified throughout the central sector of the BIIS. The criteria of Clark (1997, 1999) have been employed, allowing flow sets to be determined based on parallel conformity, length and morphology. Within the study area flow sets have been grouped into 8 regional zones, plus a series of individual valley lineations, based on related cross-cutting associations. The regional zones have been further subdivided into a series of flow phases, demarcating the relative ages of the flow sets, using the procedure outlined by Clark (1997). The higher the number the older the relative age of the flow set. Flow sets lacking cross-cutting evidence are not assigned to a specific flow phase because a relative age cannot be accurately inferred. Figure 2.3A, in the lower Vale of Eden illustrates one example of cross-cutting, with initial flow (flow phase C.4, purple arrow) moving north-westwards out into the Irish Sea Basin. This is followed by a later movement, northwards into the Solway Lowlands (flow phase C.3, red arrow) which cuts across the earlier flow. Another example, west of Stainmore Gap (Fig. 2.3B), illustrates how superimposition of landforms can be utilised in reconstructing the ice sheet dynamics. Initial flow was easterly through the Stainmore Gap (flow phase D4.3, green arrow). This was followed by a north-westerly flow (flow phase D4.1, red arrow) which partially remoulded the large drumlin positioned in the middle of diagram 2.3B, resulting in four new, superimposed drumlins. Complex flow is not just constrained to cross-cutting lineations. For example, Figure 2.2H illustrates the Holme St. Cuthbert deltaic complex overlying south-westerly

trending lineations, and the ribbed moraine nearby have been superimposed by drumlins (Fig. 2.2D) which depict a later flow phase.

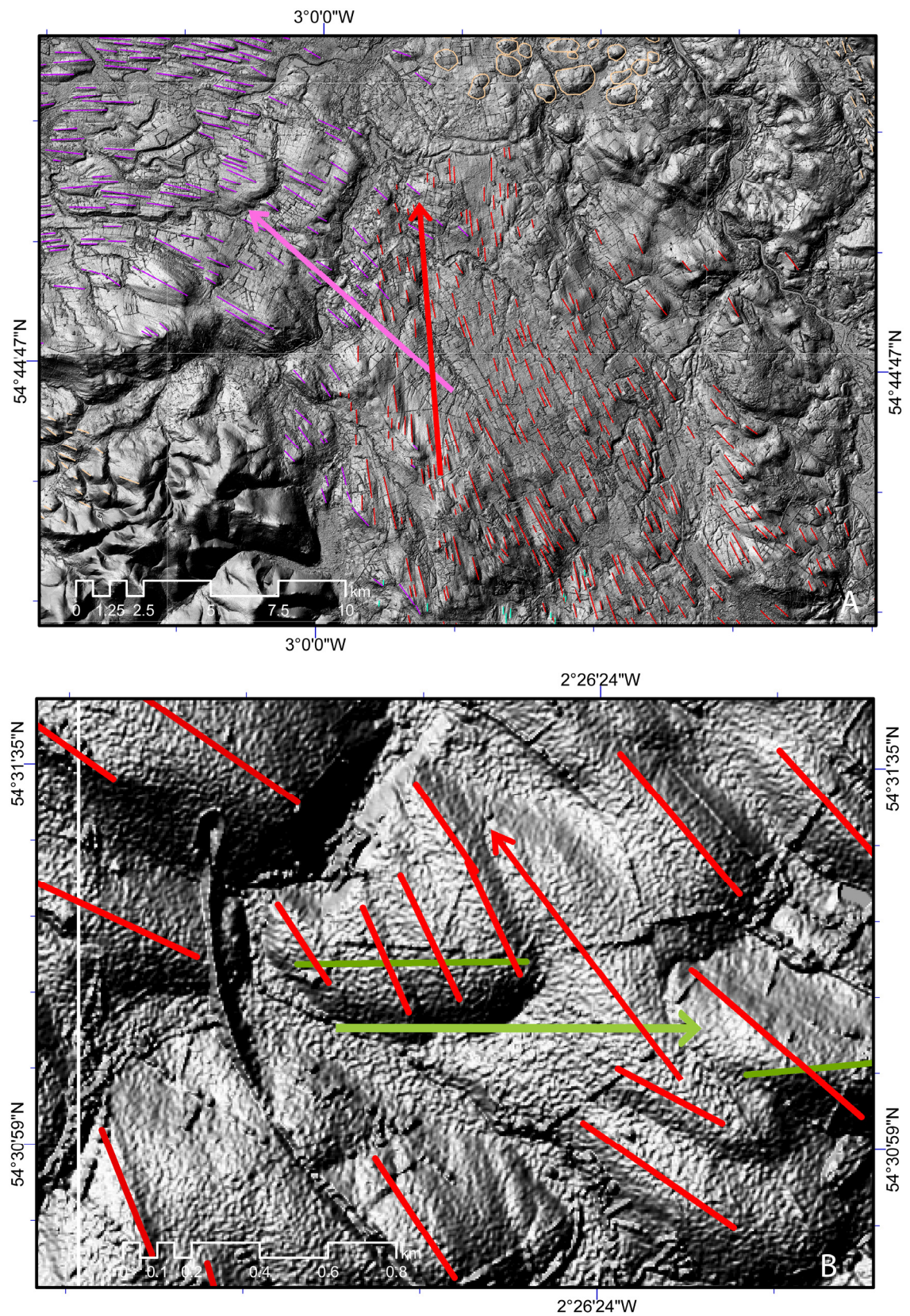


Figure 2.3: Examples of complex flow: 3A illustrates complex flow at a large scale between 2 flow sets (red and pink arrows/vectors). The earlier (pink arrow) phase was south-westerly, round the north of the Lake

< Figure 2.3 (continued)... District. This was followed by a subsequent advance (red arrow) into the Solway Lowlands. 3B demonstrates superimposed drumlin formation just to the west of the Stainmore Gap, in the Vale of Eden. An initial easterly flow across Stainmore (green arrow) was followed by a north-westerly flow (red arrow) down the Vale of Eden towards the Solway Lowlands. The west-east orientated drumlin in the centre of the image has been superimposed by four smaller drumlins. Relief shaded DEM (azimuth: 315).

#### 2.2.4 *Elongation ratios:*

Elongation ratios (length/width) provide a method for estimating the relative velocity of ice flow throughout a glacial cycle. Stokes and Clark (2001, 2002) suggest that highly attenuated subglacial bedforms characterised by high elongation ratios reflect fast ice flow. However, absolute velocities cannot be calculated as the elongation ratios are also influenced by a number of other factors including sediment supply and texture (Rose, 1987) and duration of flow (Dyke, & Morris, 1988) associated with formation. Elongation ratios across the study area fluctuate both between, and within, individual flow sets (Table 2.2). Most of the lineations have low mean elongation ratios (of 2:1 to 3:1) which suggest low flow velocities. An example of high elongation ratios is observed in the arcuate series of lineations to the north of the Lake District which sweep into the Irish Sea Basin (Fig. 2.4). The high elongation ratios (over 5:1 and as high as 12:1) relative to other flow phases in the area suggest ice was being vigorously drawn-down into the Irish Sea Basin (Eyles & McCabe, 1989; Evans & Ó Cofaigh, 2003; Ó Cofaigh & Evans, 2007; Roberts *et al.*, 2007; Thomas & Chiverrell, 2007). Recognisable differences between bedrock-moulded lineations and till-cored lineations can also be identified when elongation ratio data is draped with the surficial geology. For example, in the Tyne Gap, bedrock moulded lineations tend to be ‘stubbier’ and less elongate. The lowest elongation ratios are in the Solway Lowlands, with hummocky terrain displaying no discernable elongation characteristics (ratio of 1:1), thus suggesting that ice was very slow moving or even stagnant.

### 2.3 **Implications for BIIS flow dynamics**

The flow phases identified in the glacial geomorphological map have been used to reconstruct ice sheet dynamics within the central sector of the last British-Irish Ice Sheet during the Late Devensian (Fig. 2.5). Seven flow phases have been identified within the Lake District-Tyne Gap (LT) region. Initial ice flow phases are inferred to have occurred when ice was sufficiently thick for it to flow across the Pennines and to spill eastwards across the Tyne Gap (LT1/2/3). The initial build-up of ice in the Irish Sea Basin, predominantly driven by southerly-flowing Highland ice (e.g. Salt & Evans, 2004), would have eventually resulted in the formation of an ice dispersal centre (Eyles & McCabe, 1989; Scourse *et al.*, 1990, 1991; Greenwood & Clark, 2009), thus inhibited the

westwards movement of ice out of the central sector of the BIIS. The gradual ice-flow shift from north-east (LT1) to east (LT3) through the Tyne Gap reflects the waning influence of Lake District and Vale of Eden ice and the migration of the Irish Sea ice divide northwards into the Southern Uplands (LT4). A major flow switch, associated with the 'Blackhall Wood re-advance' (cf. Livingstone *et al.*, sub) occurred when ice drained westwards into the Irish Sea Basin (LT5) cutting-off the Tyne Gap. This was probably associated with drawdown into the Irish Sea Ice Basin (Eyles & McCabe, 1989; Ó Cofaigh & Evans, 2007; Roberts *et al.*, 2007). Superimposed on LT5 is a deltaic complex which delineates the limit of a south-easterly ice advance that crossed the Solway Firth (SF1) during or following LT6, LT7. These final two flow phases are characterised by topographically constrained flow lobes extending into the lowlands (LT6/7).

Regional zone	Flow Phase	Flow Set	Frequency	Minimum ER	Maximum ER	Mean ER	Standard Deviation
<i>A: Galloway</i>	1	1.1	177	1.49	7.07	2.96	0.97
<i>B1: Tyne Gap</i>	1	1.1	29				
	2	2.1	59	1.19	3.53	2.18	0.56
		2.2	220	1.12	4.69	2.21	0.68
	3	3.1	609	1	7.76	2.52	0.94
		3.2	126	1.01	4.23	2.01	0.68
		3.3	26	1.2	2.82	1.92	0.38
	4	4.1	236	1.34	7.26	3.02	1.23
		4.2	472	1	7.33	2.4	0.99
		4.3	149	1.04	11.95	4.36	2.18
	5	5.1	329	1	9.91	2.72	1.33
	6	6.1	309	1.01	10.21	2.19	0.85
		6.2	161	1	9	2.29	1.12
		6.3	275	1.36	11.98	3.5	1.61
	4-6	7.1	19	1.23	4.99	2.66	0.96
	1-6	8.1	43				
<i>B2: Tyne Gap</i>	1/2	1.1	32	2.11	5.66	3.56	0.85
	1/2	1.2	60	1.11	5.45	2.44	0.86
<i>C: Vale of Eden</i>	1	1.1	116	1.21	4.84	2.17	0.6
	2	2.1	64	1.19	5.9	2.35	0.87
	3	3.1	653	1	6.78	2.94	1.01
		3.2	243	1.01	4.46	2.22	0.58
	4	4.1	573	1	11.73	3.23	1.61
		4.2	50	1.42	3.84	2.15	0.58
	5	5.1	38	1.51	3.57	2.48	0.52
		5.2	57	1.05	3.95	2.02	0.59
	1-2	6.1	48	1.04	2.91	1.9	0.42
		6.2	72	0.49	4.76	1.92	0.58
	1-4	7.1	13	1	5	2.69	0.99
	1-5	8.1	77	1	4.02	2.06	0.68
<i>D1: Stainmore</i>	1	1.1	1	1	3.11	1.77	0.42
	2	2.1	134	1.03	3.95	1.96	0.53
	3	3.1	32	1.16	2.85	1.95	0.46
<i>D2: Stainmore</i>	1	1.1	84	1.25	4.46	2.24	0.69
<i>D3: Stainmore</i>	1	1.1	122	1.15	4.29	2.19	0.52
		1.2	25	1.28	5.47	2.53	0.79
	2	2.1	141	1.28	1.88	1.49	0.18
	3	3.1	25	1.36	2.91	1.83	0.34
<i>D4: Stainmore</i>	1	1.1	368	1.06	5.68	2.4	0.76
	2	2.1	406	1	9.25	2.42	0.94
	3	3.1	67	1.52	6.08	2.98	0.87
		3.2	85	1.22	5.46	2.45	0.69
		3.3	30	1.38	6.38	3.01	1.15
	2-3	4.1	21	1.24	3.95	2.24	0.71
	2-3	5.1	16	1.54	3.21	2.3	0.46
		5.2	13	1.78	4.88	2.86	0.76
		5.3	24	1.49	3.42	2.32	0.52

Table 2.2: Elongation ratios: statistics



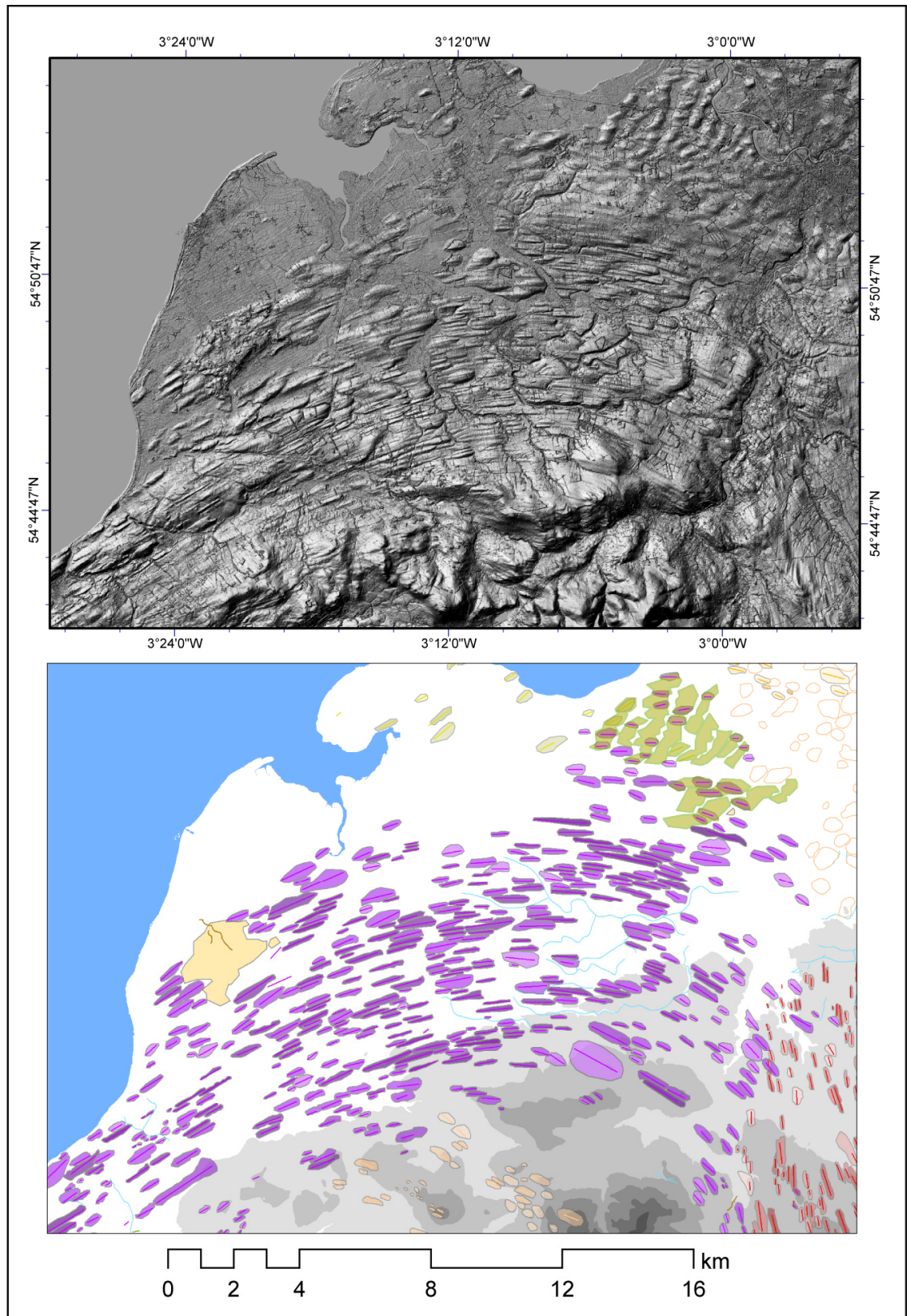


Figure 2.4: Ice streaming round the northern Lake District into the Irish Sea Basin. Relief shaded DEM (azimuth: 315).

Four flow phases have been recognised within the Stainmore Gap (ST) region, characterised by an increasingly prevalent Tees Valley ice flow forcing ice south-eastwards towards the Vale of York (ST3/4). The Stainmore group are likely to have been formed relatively early in the Late Devensian based on the cross-cutting of flow set ST1 by LT6. Ice must have been sufficiently thick at ST1 for ice sourced in the Howgill Fells and Lake District to overtop the Stainmore col. Ice build-up probably resulted in an ice divide developing in the Vale of Eden between ice converging on both Stainmore and the Tyne Gap. The flow in the Vale of Eden (ES1) is difficult to interpret as despite showing no conceivable difference with LT6, it veers eastwards into the Stainmore Gap. Stoss and lee forms have also been interpreted as indicative of southerly flow (Hollingworth, 1931), although on the NEXTMap data this is hard to distinguish.

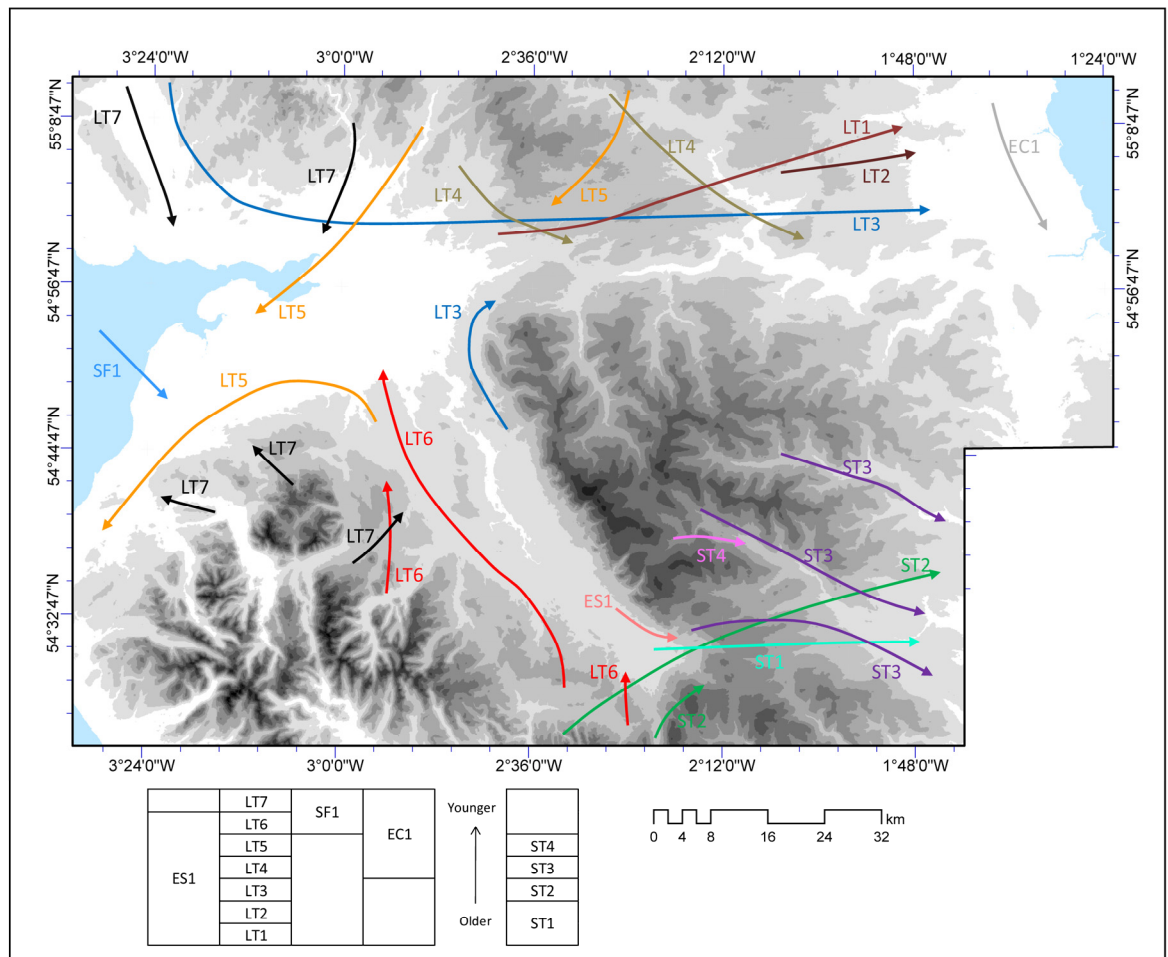


Figure 2.5: Generalised flow phases of ice dynamics in the central sector of the last BIIS during the Late Devensian and Table showing relative chronology flow stacks from cross-cutting bedform relationships. Flowsets are stacked vertically according to cross-cutting relationships (with the bottom block being the oldest) with horizontal blocks denoting discrete regional groupings.

Ice moving southwards down the east coast (EC1) is unrelated to other flow sets within the area. However, the absence of any easterly orientated lineations within this region suggests that EC1 occurred after LT4, wiping away or covering evidence of the Tyne Gap ice flow.

While the lowland regions of the central sector of the BIIS show evidence for extensive glaciation and landform generation indicative of warm-based ice flow, within the upland regions (> 400 m O.D.) of the Lake District and the northern Pennines glacial bedform features are much rarer. This has particular resonance in relation to the basal thermal regime of the field area (Kleman & Glasser, 2007). The interfluves which characterise the northern Pennines suggest preservation of a previous drainage network by cold-based ice (Kleman & Glasser, 2007). Indeed Mitchell (2007) proposes a model of temperate ice-flow through major valley networks including Weardale, the Allendales, the South Tyne and Teesdale based on subglacial bedform evidence and erratics (e.g. Vincent, 1969), juxtaposed with a cold-bedded ice plateau on the higher non-drumlinised ground.

## **2.4 Conclusions**

In summary, the presented geomorphological map (see back of thesis for A3 pull-out) illustrates the wide variety and complexity of landform deposition during the Main Late Devensian in the central sector of the British-Irish Ice Sheet. A palimpsest geomorphic signature characterised by cross-cutting glacial landforms is interpreted as evidence for dynamic ice flow in response to migrating ice dispersal centres and ice divides. The variety of landforms details the evolving behaviour of the ice sheet through periods of both advance and decay. Based on this data, a relative chronology has been established which begins to reconstruct the glacial history throughout the Main Late Devensian, thus providing a framework for future analysis and interpretation.

## **2.5 Software**

The NEXTMap data was processed and digitised using ArcGIS 9.1 (ArcMap and ArcCatalogue). Adobe Illustrator CS was used in the production of the map.

## **2.6 Acknowledgments**

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## Chapter 3

# The palaeoglaciology of the central sector of the British and Irish Ice Sheet: reconciling glacial geomorphology and preliminary ice sheet modelling

Evans, D.J.A., Livingstone, S.J., Vieli, A. & Ó Cofaigh C. 2009. The palaeoglaciology of the central sector of the British and Irish Ice Sheet: reconciling glacial geomorphology and preliminary ice sheet modelling. *Quaternary Science Reviews*, 28; 735-757.

## Abstract

Digital elevation models of the area around the Solway Lowlands reveal complex subglacial bedform imprints relating the central sector of the LGM British and Irish Ice Sheet. Drumlin and lineation mapping in four case studies show that glacier flow directions switched significantly through time. These are summarized in four major flow phases in the region: Phase I flow was from a dominant Scottish dispersal centre, which transported Criffel granite erratics to the Eden Valley and forced Lake District ice eastwards over the Pennines at Stainmore; Phase II involved easterly flow of Lake District and Scottish ice through the Tyne Gap and Stainmore Gap with an ice divide located over the Solway Firth; Phase III was a dominant westerly flow from upland dispersal centres into the Solway Lowlands and along the Solway Firth due to draw-down of ice into the Irish Sea Basin; Phase IV was characterised by unconstrained advance of Scottish ice across the Solway Firth. Forcing of a numerical model of ice sheet inception and decay by the Greenland ice core record facilitates an assessment of the potential for rapid ice flow directional switching during one glacial cycle. The model indicates that, after fluctuations of smaller radially flowing ice caps prior to 30 ky BP, the ice sheet grows to produce an elongate, triangular-shaped dome over NW England and SW Scotland at the LGM at 19.5 ky BP. Recession after 18.5 ky BP displays a complex pattern of significant ice flow directional switches over relatively short timescales, complementing the geomorphologically-based assessments of palaeo-ice dynamics. The palaeoglaciological implications of this combined geomorphic and modeling approach are that: (a) the central sector of the BIIS was as a major dispersal centre for only *ca.* 2.5 ka after the LGM; (b) the ice sheet had no real steady state and comprised constantly migrating dispersal centres and ice divides; (c) subglacial streamlining of flow sets was completed over short phases of fast flow activity, with some flow reversals taking place in less than 300 years.

## 3.1 Introduction

Recent modelling of the dynamics of the British and Irish Ice Sheet (BIIS) through the last glacial cycle has identified major dispersal centres and ice flow pathways (Boulton & Hagdorn, 2006). Due to their spatial resolution, such models often fail to reconcile the temporal and spatial complexities of the glacial geomorphological evidence that documents regional ice sheet build-up and decay. It is this complex geomorphic signature, particularly the subglacial bedform record, which provides us with an increasingly clear impression of the multiplicity of flow phases that characterised ice sheet evolution during the Last Glacial Maximum (LGM) (Punkari, 1982, 1993; Boulton & Clark, 1990a, b; Clark, 1997; Clark *et al.*, 2000; Clark & Meehan, 2001). Although glacial bedform mapping in North America and Scandinavia has yielded significant advances in understanding the dynamic nature of ice sheet behaviour, the subglacial features of the British Isles have only recently been analysed systematically in areas of formerly complex ice sheet flow (e.g. Salt & Evans, 2004). As a result numerical models of the BIIS (e.g. Boulton & Hagdorn, 2006) do not focus on flow directional changes associated with multiple ice dispersal centres but rather give the impression that ice flowed from relatively static domes such as that over western Scotland.

We present initial geomorphological mapping and numerical modeling results from a project aiming to reconstruct the palaeoglaciology of the former centre of the BIIS. This focuses on the location of palaeo-ice divides, dispersal centres, flow trajectories and flow phasing through the mapping of subglacial bedforms and the identification of their cross-cutting relationships. Forcing of a numerical model of ice sheet inception and decay by the Greenland ice core record allows us to assess the potential for rapid ice flow directional switching during one glacial cycle. The ice-flow trajectories reproduced by the model, while not exactly replicating the geomorphologically constrained flow phases, demonstrates the highly dynamic nature of ice sheet flow patterns, and emphasizes the short periods of time over which significant flow switches, some involving ice flow reversal, can take place at the centre of an ice sheet. Future applications of our data include usage in the testing of spatial and temporal variability of linkages between ice sheet behaviour, ocean/climate change, sea level rise and internal driving mechanisms (e.g. binge-purge cycles, topographic influences etc.), especially when more complex boundary conditions (e.g. sea level rise) are incorporated into numerical models.

### **3.2 Previous palaeoglaciological reconstructions**

The central sector of the former BIIS (Fig. 3.1) was a complex multi-sourced region of competing ice flows draining the Southern Uplands, Pennines, Lake District and the Irish Sea, resulting in the production of a complex geomorphic signature of flow dynamics. Early research recognised three phases of ice flow into the region, including an early “Scottish advance”, a later “Main glaciation” and then a re-advance phase (Trotter & Hollingworth, 1932). The early Scottish advance is

recorded by a stratigraphically older reddish-brown till containing Scottish erratics and documenting Scottish ice advance into the Tyne Gap and up Edenside as far as Stainmore, diverting Lake District ice over the Stainmore Gap (Trotter, 1929; Hollingworth, 1931; Huddart, 1970, 1971). The easterly ice-flow direction across Stainmore is recorded by the unique Shap granite, Dufton Pike granite and Permian Brockram dispersal train (Trotter, 1929; Hollingworth, 1931). Easterly ice flow through the cols of Stainmore (533 m) and Cold Fell (625 m) required significant ice sheet thickening over the Southern Uplands and Lake District, whereas flow along the lower Tyne Gap (152 m) appears to have been maintained throughout most of the last glaciation. Easterly flowing ice over the higher terrain to the east of Edenside (i.e. Cross Fell and surrounding summits) generally indicates that the region became a significant dispersal centre (Hollingworth 1931).

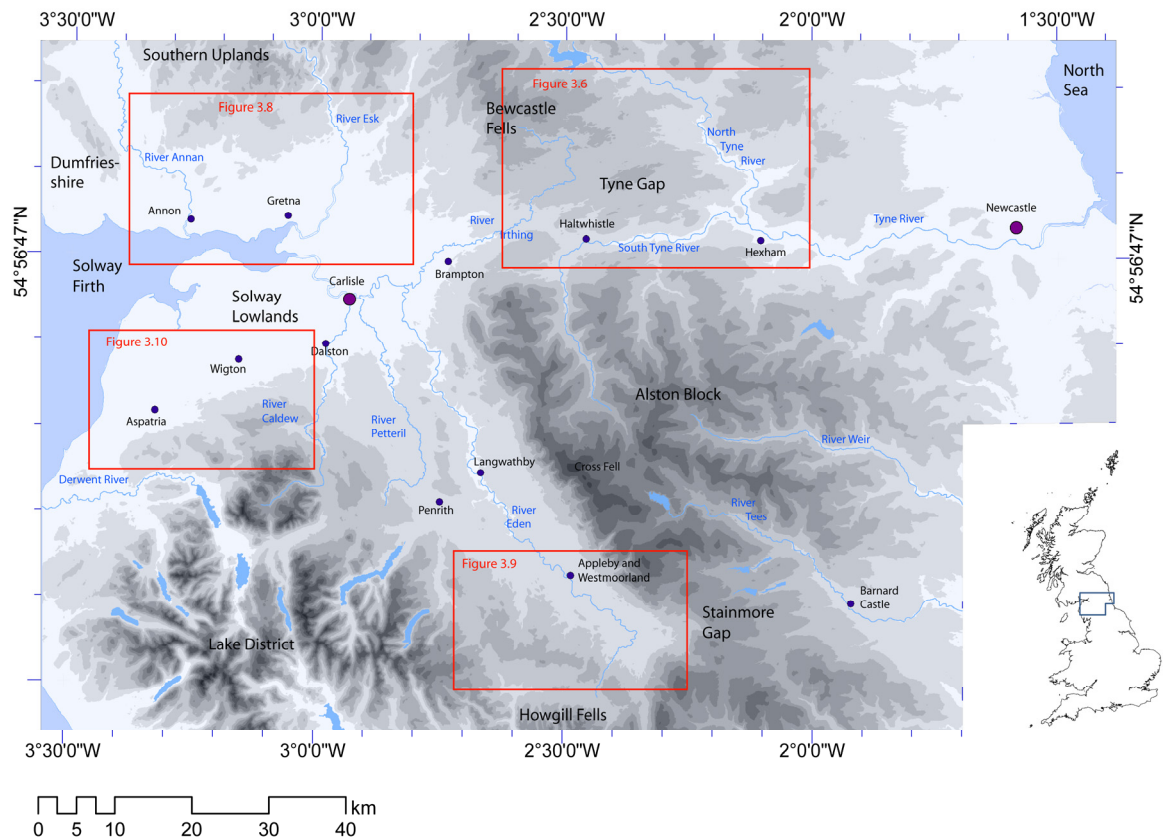


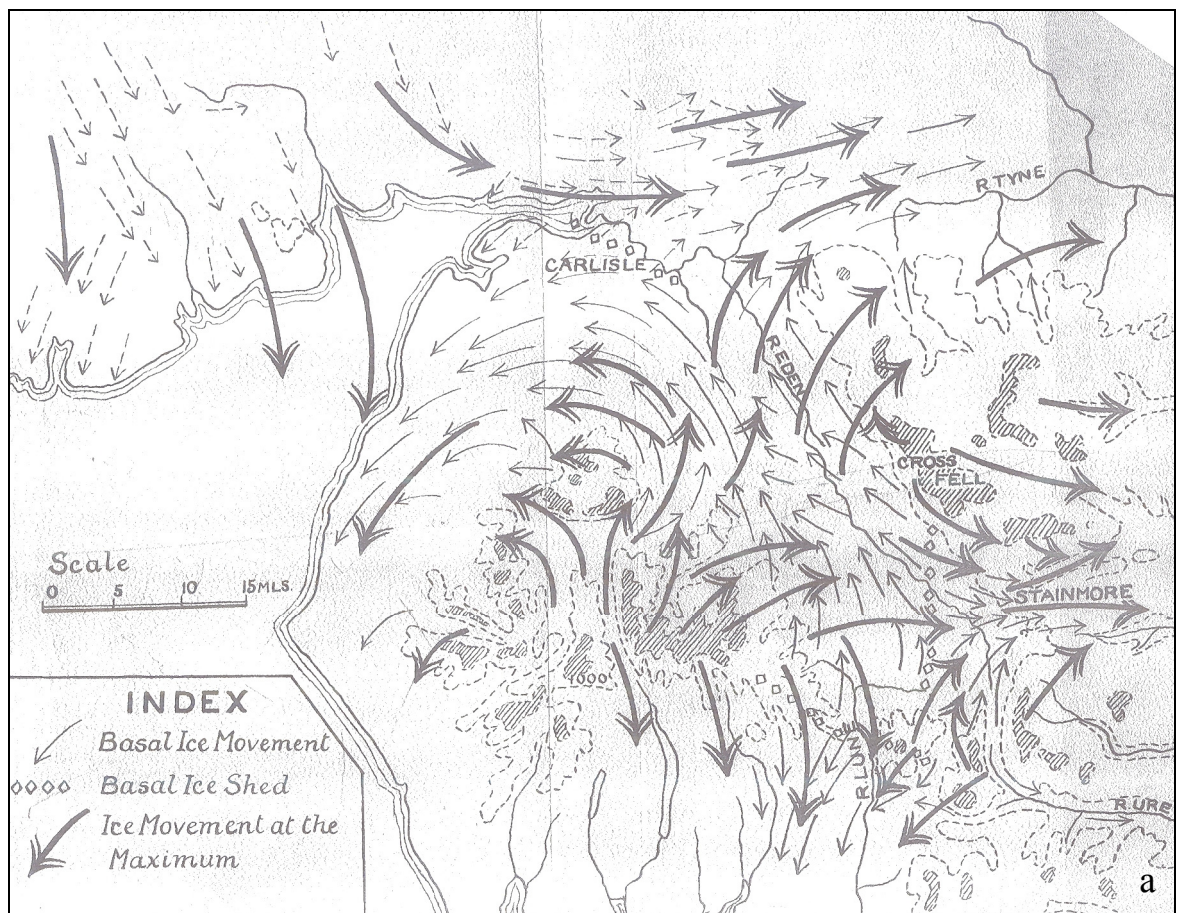
Figure 3.1: Location map of the central sector of the BIIS with areas covered by case studies located by boxes and names of sites mentioned in text. The research area includes the Southern Uplands, Lake District, Alston Block, Vale of Eden, Solway Lowlands, Tyne Gap and Stainmore Gap.

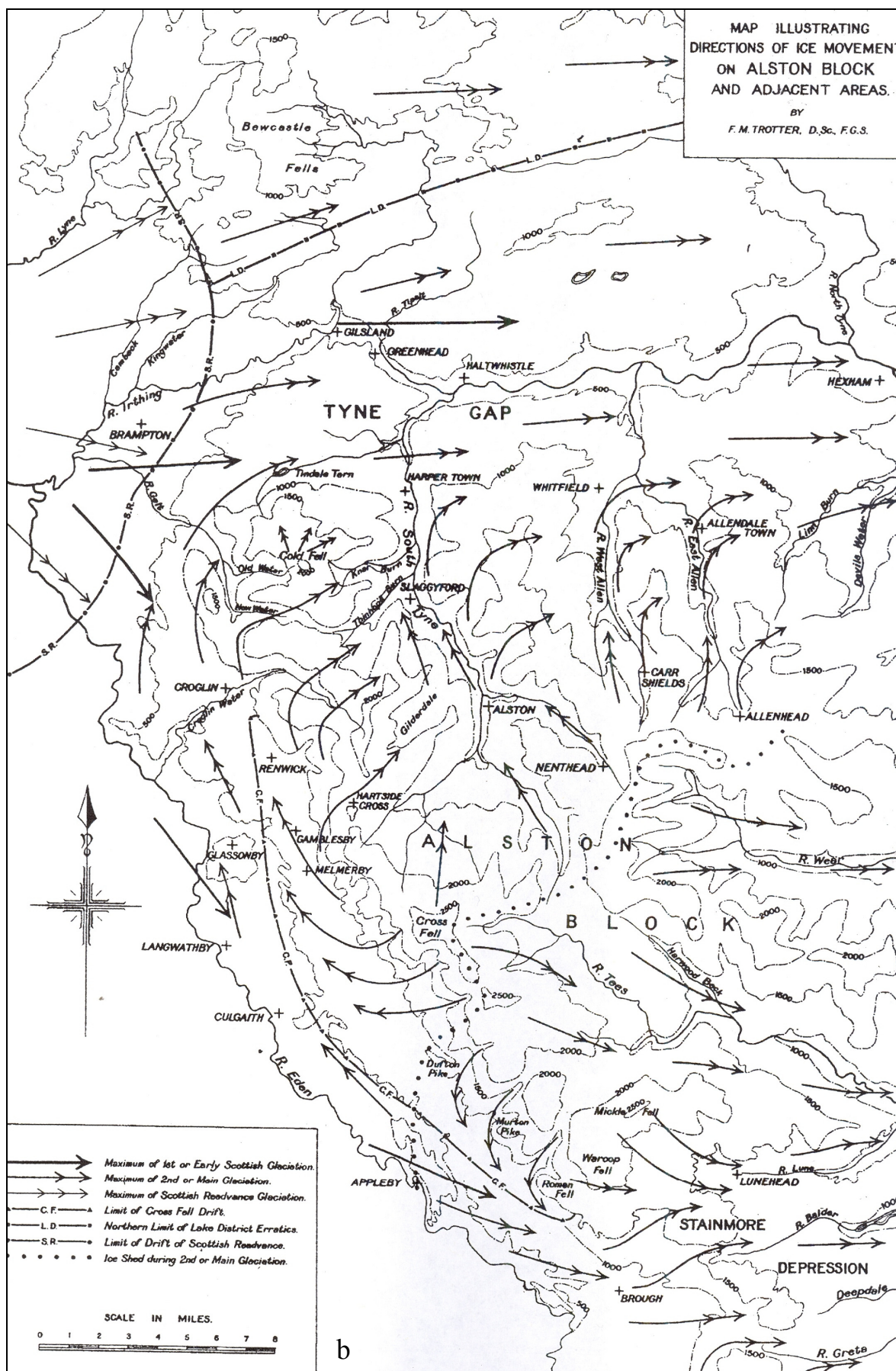
The southerly flow of Scottish ice is reported to have been superseded by a northerly ice-flow during the “Main Glaciation”. Laminated clays at Langwathby (Fig. 3.2; Goodchild, 1875) between the deposits of the ‘Scottish’ and ‘Main’ glaciations suggest at least partial deglaciation of the area, although the magnitude of retreat is uncertain (Hollingworth, 1931). Lake District ice flowing



northwards down the Vale of Eden is recorded stratigraphically by an extensive red, sandy till containing Borrowdale volcanics and Penrith sandstone. The northerly, down valley flow is indicated by the change in content of these erratics from 90% in the south to 85% in the north (Trotter, 1929). Additionally, drumlin orientations to the north of the Lake District were interpreted by Trotter (1929) and Hollingworth (1931) as evidence of Lake District ice flow into the Solway Lowlands, where it was deflected west towards the Irish Sea because of “congestion” with Scottish ice to the north (Fig. 3.2).

A corridor of prominent streamlining occurs through the Tyne Gap (Johnson, 1952), where bedform lineation and elongation has been interpreted as the imprint of a former ice stream (Bouledrou, *et al.*, 1988; Everest *et al.*, 2005). Tributary ice flows from the south, west and north can also be identified in the bedform record but the sedimentary boundary between the Scottish and Lake District ice in the Tyne Gap is indistinct, and characterised by an increase in Lake District erratics (Borrowdale volcanics, Carrock Fell gabbro, Threlkeld microgranite and Penrith sandstone) towards the south (Trotter, 1929). Along its length from west to east the Tyne Gap is increasingly dominated by blue-grey, Scottish sourced till (Trotter, 1929), the drop off in Lake District erratics being interpreted by Derryhouse (1902) as a product of the influence of Scottish ice flowing south down the North Tyne.







< Figure 3.2: Early interpretations of glacier flow directions in the central BIIS based on: (a) Trotter (1929); and (b) Hollingworth (1931).

The development of a separate Pennine icefield/dispersal centre during glaciation has been promoted by Dakyns *et al.*, (1891), Derryhouse (1902), Vincent (1969) and Mitchell (2007) based upon the occurrence of a distinctive local till at Cross Fell, which documents flow from the Cross Fell plateau and down Edenside to coalesce with Lake District ice feeding into the Tyne Gap (Trotter, 1929). This ice flow direction is difficult to reconcile with contemporary Eden ice flowing southward and over Stainmore and this, together with complex drumlin orientations in the Eden valley, compounded early attempts to reconstruct palaeo-ice flows (Hollingworth, 1931; Clark, R. 2002). For example, around Appleby, Hollingworth (1931) recognised drumlins with strikingly different flow directions (Fig. 3.2); west of Appleby movement was initially eastwards with an overprinted flow down Edenside, and east of Appleby there was an alignment that suggested ice flow towards Stainmore (Clark, R. 2002). Hollingworth (1931) proposed a “basal ice shed” in order to explain ice flowing in two different directions (Figure 3.2, lower). This is now known to be glaciological implausible, and Rose and Letzer (1977) were the first to suggest that the region contained overprinted subglacial bedforms that record shifting ice divides and migrating ice dispersal centres during a single glacial episode. The mobility of the local Pennine ice divide has similarly been charted by Mitchell (2007) based upon superimposed drumlins.

A further subglacial bedform enigma occurs in the Solway Lowlands, into which ice flowed from the Scottish Southern Uplands and from the Lake District. Erratic trains, particularly from distinct granite outcrops around the Galloway area, indicate that ice radiated out from a Southern Upland dispersal centre and flowed down valleys such as the River Nith (Tipping, 1999). Grey-blue till, free of Lake District erratics, crops out in river sections of the Lyne, Irthing and Tipalt, marking the southern limit of Scottish ice at the southern edge of Bewcastle Fells. Subglacial bedforms appeared to indicate to early workers that the build-up of ice in Edenside forced Scottish ice eastwards but itself flowed west (Fig. 3.2; Trotter, 1929), resulting in a palaeoglaciological reconstruction with the implausible scenario of contemporary ice flows in opposite directions. Recent developments in understanding spatial and temporal patterns in subglacial bedforms indicate that the enigmatic drumlins of NW England and SW Scotland are best understood when viewed as a landform palimpsest in which bedforms are overprinted (Mitchell & Clark, 1994).

The concept of subglacial bedform overprinting is now widely accepted in the research field of palaeoglaciology and has been successfully applied to reconstructions of the Laurentide and Fennoscandinavian Ice Sheets (e.g. Boulton & Clark, 1990a, b; Clark 1993, 1994, 1997; Kleman 1994; Kleman *et al.* 2006). Complex subglacial bedform palimpsests have been recognized in the British Isles in a number of localities, for example by Rose and Letzer (1977) in the Vale of Eden, Clark and Meehan (2001) in Ireland and by Salt and Evans (2004) for the Southern Uplands ice

dispersal centre where the subglacial bedforms are of multiple ages and have been superimposed or partially removed (Fig. 3.3). Recently, with the development of GIS techniques, mapping has extended to the ice-sheet scale, with Greenwood & Clark (2008, 2009a,b) producing a glacial geomorphological map and associated model of ice sheet growth and decay for the Irish Ice Sheet, and Hughes (2008) producing a glacial geomorphological map and reconstruction of the British Ice Sheet from the NEXTMap DEM. Salt and Evans (2004) provide evidence of four early, topographically unconstrained ice flow stages (A-D) and three late topographically constrained stages (E-G) for SW Scotland. It was the flows from the Southern Upland dispersal centre that impinged upon the Solway/Eden Valley area and these, from oldest to youngest, were: Stage A flow towards the southwest; pre-Stage B southerly flow; Stage B southwesterly flow; Stage C south-southwesterly flow; and Stages F and G when ice flowed down major valleys such as the Cree and Ken and along Loch Ryan to terminate somewhere in the Solway Firth. In addition, some southeasterly regional flows of unknown relative age are apparent on the east side of the Southern Uplands (New Galloway district in Fig. 3.3). In NW England, overprinted drumlins have been used to verify the reversal of ice flow during the last glaciation in the Vale of Eden (Mitchell & Riley, 2006) and to reconstruct ice divide migration over the western Pennines and Howgill Fells (Mitchell, 1994). The potential for identifying further subglacial bedform complexity in the British Isles and employing it in palaeoglaciological reconstructions of the BIIS has been reviewed by Clark *et al.* (2006).

Because it contains a number of former upland ice dispersal centres that coalesced to form the core of the BIIS during the LGM, NW England potentially contains a complex record of flow phasing in its glacial landform-sediment assemblages (cf. Hollingworth, 1931; Trotter & Hollingworth, 1932). Salt and Evans (2004) identified superimposed multiple flow events to the north where Southern Upland ice competed with flow from the Scottish Highlands (Fig. 3.3), and there is no reason to believe that similar ice sheet flow complexity is not manifest in the subglacial bedform record over NW England, in part linked to the dynamics of ice over SW Scotland. This has been acknowledged by Merritt and Auton (2000) in their reconstruction of ice flow events along the Cumbria coast. They use a lithostratigraphic approach to identify oscillations of Scottish/Irish Sea ice and Lake District ice that were significant enough to allow the development of unglaciated enclaves and glacial lakes along the Cumbria coast through the later stages of the last glacial cycle. Significantly, two major re-advances have been identified by Merritt and Auton (2000) at *ca.* 17 <sup>14</sup>C ka BP or *ca.* 19.5 cal ka BP (Gosforth Oscillation) and *ca.* 14 <sup>14</sup>C ka BP or *ca.* 16.8 cal ka BP (Scottish Re-advance), although the chronological control on these events is equivocal.

More secure is the chronology on ice sheet maximum limits and re-advances in the Irish Sea Basin. Scourse (1991), Scourse and Furze (2001), Hiemstra *et al.* (2005) and Ó Cofaigh and Evans (2007) provide evidence that the BIIS reached the Scilly Isles and the south coast of Ireland after 20.2 <sup>14</sup>C ka BP (23.9 cal ka BP). To the north, McCabe and Clark (1998, 2003) and McCabe *et al.* (1998,

2005, 2007) have identified two subsequent re-advances, the Clogher Head re-advance between 15 and 14.2  $^{14}\text{C}$  ka BP (18.3-17.0 cal ka BP) and the Killard Point re-advance after 14.2  $^{14}\text{C}$  ka BP (17.0 cal ka BP). The latter was a pan-Irish Sea response by the BIIS to Heinrich Event 1, which involved a re-advance to the Bride Moraine on the Isle of Man. The northern plain of the Isle of Man contains evidence for repeated marginal oscillation during retreat, with the chronology for ice contact sediments of the Jurby and older Orrisdale Formations constrained by overlying organic materials (Thomas *et al.*, 2004). Specifically, terrestrial plant remains from the basal layers of kettle basins date to the beginning of the late glacial interstadial at 14.0-14.4 and 14.2-15.0 cal. kyrs BP (Roberts *et al.*, 2007). Cosmogenic nuclide dates on the Wester Ross Re-advance moraine indicate that the ice sheet was onshore in NW Scotland during Heinrich event 1 (Everest *et al.* 2006). Although some local glacial stratigraphies clearly equate ice marginal re-advances with global climate signals such as Heinrich events (McCabe 1996; McCabe *et al.* 1998), evidence for re-advances is widespread around the Irish Sea Basin and many cannot be explained by external climate drivers (e.g. Thomas & Summers, 1983; Evans & O Cofaigh, 2003; Thomas *et al.*, 2007).

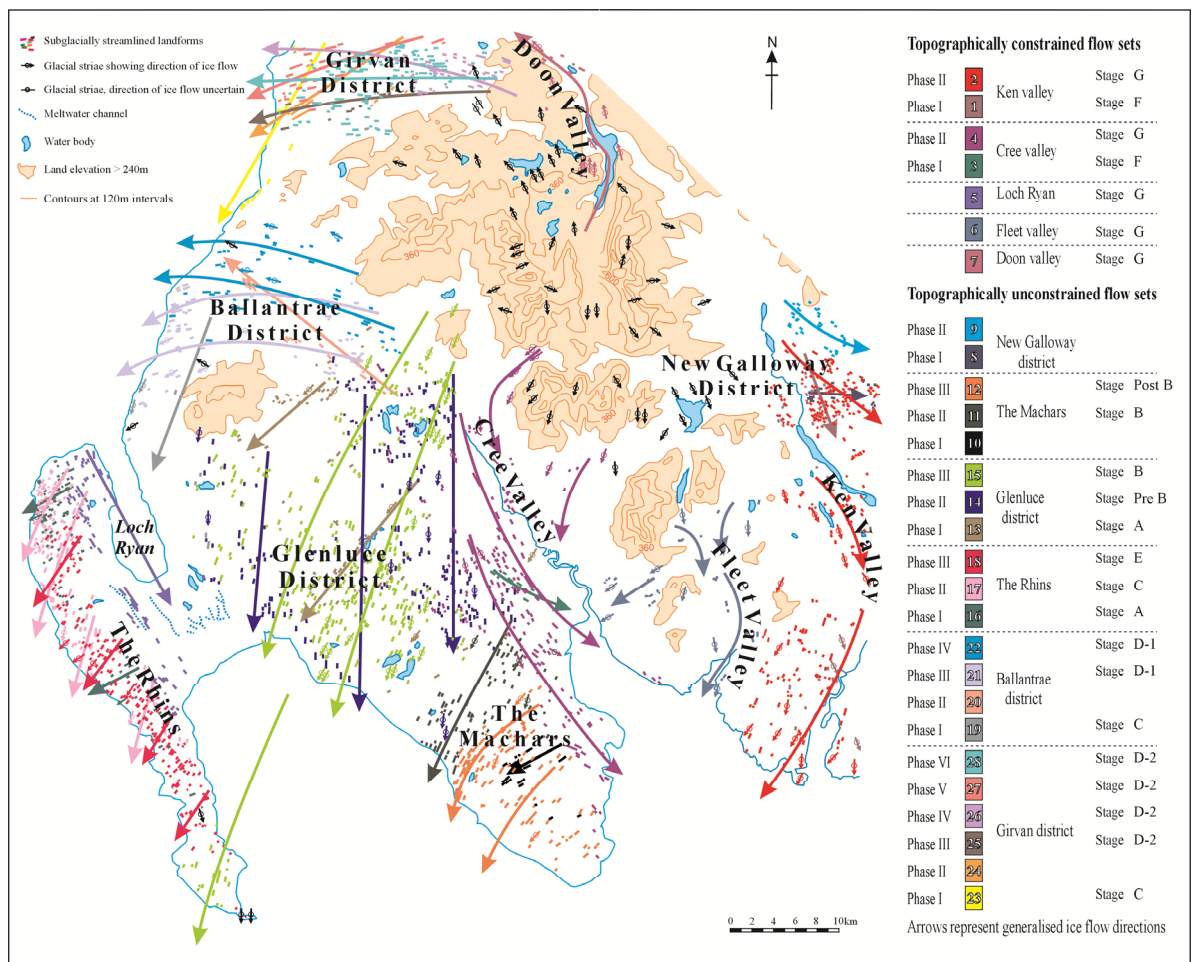


Figure 3.3: Map of cross-cutting flow sets in SW Scotland (from Salt & Evans 2004).

### 3.3 Rationale and approach

Despite the success of McCabe and Clark (1998), McCabe *et al.* (1998, 1999, 2005) and P.U. Clark *et al.* (2004) in correlating palaeo-ice sheet marginal oscillations in NE Ireland with North Atlantic and even global climate-ocean events, the behaviour of the BIIS throughout the last glacial cycle is poorly understood, due to patchy and largely non-systematic mapping of glacial landforms (see C.D. Clark *et al.*, 2004 and Evans *et al.*, 2005 for a review) and a weakly constrained chronological control on glacial deposits/events. Although there have been some attempts to reconstruct the palaeoglaciology of the last BIIS (Boulton *et al.*, 1977, 1985), these reconstructions have been based on either non-systematic mapping of spatially limited data, or inverse approaches using sea level histories (e.g. Shennan *et al.* 2002). Moreover, as McCabe *et al.* (2005) point out, these models are “static” because they are based on subglacial flowline indicators (bedforms) of mixed ages.

Some zones of former fast flow or ice streaming have been identified based on geomorphic criteria (Stokes & Clark 1999, 2001; Clark *et al.*, 2004) and sedimentological evidence (e.g. Merritt *et al.*, 1995; Evans & Ó Cofaigh 2003). It is also becoming clear that several generations of ice flow lineations exist and demonstrably cross-cut each other (e.g. Clark & Meehan, 2001; Salt & Evans, 2004). Despite some systematic attempts to map regional subglacial bedforms (e.g. Knight & McCabe, 1997; McCabe *et al.*, 1998, 1999; Clark & Meehan, 2001; Salt & Evans, 2004; Clark *et al.*, 2006) our knowledge of BIIS palaeo-ice flow dynamics and their relationship to changes in ice sheet configuration and size remains at a relatively crude level compared to advances in research on the Laurentide and Scandinavian Ice Sheets. Although the BIIS has had a long history of research, previous work has predominantly concentrated on sediment stratigraphy and correlation, at the expense of investigations into spatial and temporal changes in ice-sheet dynamics. In order to resolve this and to elucidate BIIS interaction with ocean-atmosphere events over the last glacial cycle, systematic subglacial bedform and sediment mapping, constrained by robust ice-marginal chronologies, is crucial. This paper reports our attempts to identify and assess subglacial bedform overprinting at the former centre of the BIIS during the last glacial cycle with respect to the location of palaeo-ice divides, dispersal centres, flow trajectories and flow phasing. We specifically address the interactions between shifting ice masses in the three critical areas of ice build-up and dispersal (Southern Uplands, Lake District, northern Pennines) and establish their changing flow patterns and phasing. This has been achieved through reconciling the results of: (a) the mapping of subglacial landforms; and (b) the development of a numerically driven ice sheet model that depicts changing ice flow phases through the last glacial cycle; the model also provides a chronological framework for ice flow switching, because it is driven by the Greenland ice core record.

Subglacial bedform mapping involves the on-screen digitization of linear/streamlined landforms from NEXTMap Britain digital elevation model (DEM) data. This is a 5 m spatial resolution DEM derived using airborne interferometric synthetic aperture radar. Regional ice sheet flow phases are

reconstructed through the identification of flow sets or assemblages of subglacially streamlined landforms that record a dominant ice flow direction. We employed the criteria of Clark (1997, 1999) and Salt & Evans (2004) to identify ice flow tracks/fans and palaeo-ice streams and determine temporal sequencing based upon overprinting. Based on the identification of their parallel conformity, length and morphology, glacial lineaments can be divided into “flow sets” (Fig. 3.4). A “flow set” can be defined as a collection of glacial features formed during the same flow phase and under the same conditions (Clark, 1999). Flow phase sequencing constitutes a relative chronology of ice-flow events, because older flow sets are separated in time by those that overprint them. These superimposed or cross-cutting subglacial landforms survive because of the ability of ice sheets to preserve geomorphological flow features (Kleman, 1994). The co-existence of landforms superimposed at different orientations represents a palimpsest of different ice-flow events and, therefore, a series of chronologically distinct relative flow phases (Boulton and Clark, 1990; Clark, 1993). Several assumptions form the basis for much of the interpretation (Kleman and Borgstrom, 1996): (i) basal sliding requires a thawed bed; (ii) lineations can only form if basal sliding occurs; (iii) lineations are created in alignment to the local flow and perpendicular to the ice-surface contours at the time of creation; (iv) frozen bed conditions inhibit re-arrangement of the subglacial landscape. In terms of ice flow evolution, cross cutting lineations can be interpreted in terms of migrating ice divides, ice stream activation and lobate retreat (Clark, 1997) all of which leave unique signatures (Fig. 3.5).

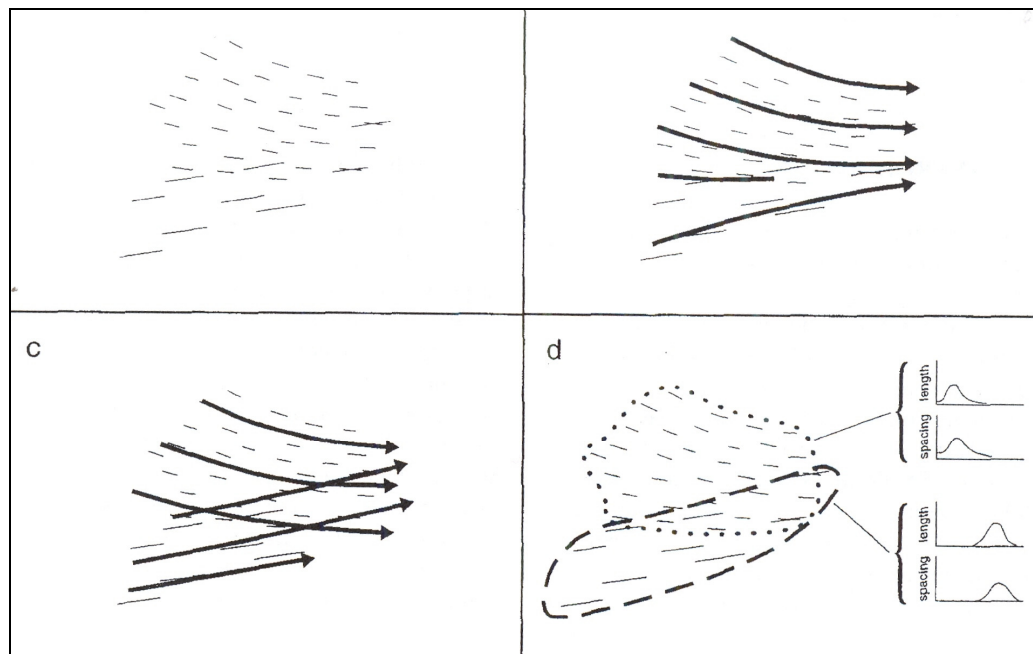


Figure 3.4: Schematic diagram to show the procedures employed in the grouping of subglacial flow sets: (a) the raw data showing lineation pattern of two separate sets of bedforms; (b) example of outdated attempts to interpret flow direction from lineations by blending the bedforms of two separate and overprinted ice flows; (c) example of more recent approach to separate the bedforms and interpret them as the products of two flow sets; (d) morphological criteria for flow set characterisation (from Clark 1997).

Smith *et al.* (2001, 2005) identified a series of problems with mapping glacial landforms from remote sensing datasets, including azimuth biasing, relative size of bedforms and landform signal strength (the degree to which individual landforms can be distinguished from other features). The glacial bedforms being mapped in this study are all distinctive landforms that are much larger than 5 m (the limit of the data resolution). The effect of light biasing has been lessened by repeated mapping from two orthogonal light directions (315° and 45°) and a ‘bird’s eye’ view which reveals slope curvature by shading and flat areas as light (Smith *et al.*, 2005).

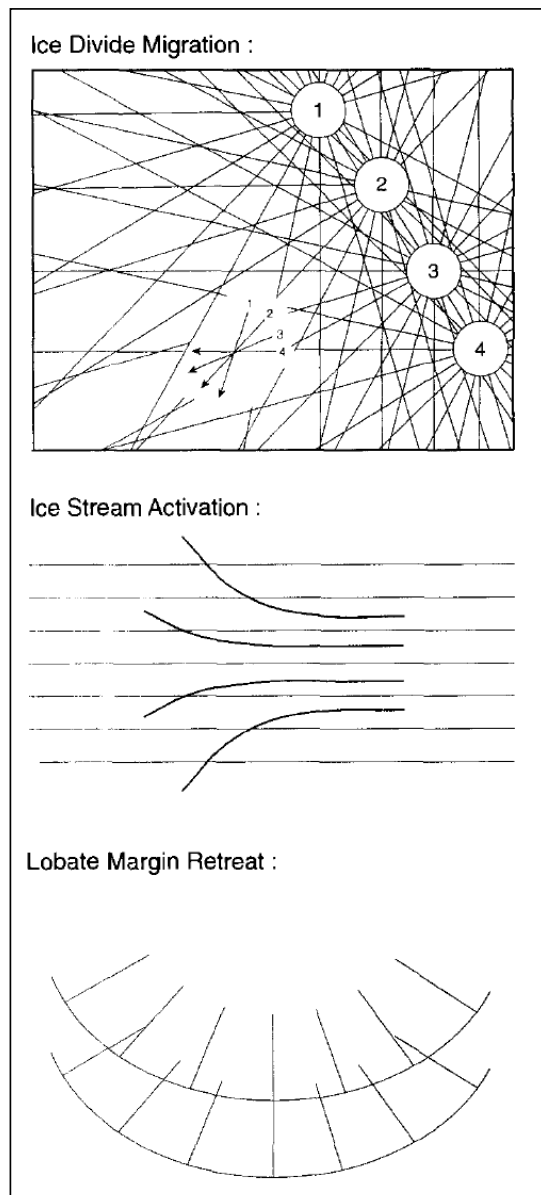


Figure 3.5: Three conceptual models of subglacial bedform overprinting (from Clark 1997). Upper diagram shows the impact of ice divide migrations through time, producing four cross-cutting flow sets at one example location. Middle diagram shows the imprints of ice stream flow, in trunk or onset zones. Lower diagram shows the impact of lobate margin recession.

Table 3.1 illustrates the glacial landforms which have been identified and mapped within the central sector of the BIIS.

<b>Landform type</b>	<b>Number mapped</b>
Subglacial lineations	8050
Hummocky moraine	102
Ribbed (Rogen) moraine	18
Meltwater channels	800
Eskers	13
Glaciofluvial sediment accumulations	36
Transverse ridges	3

Table 3.1: Mapped glacial landforms

Glacial landforms were mapped manually by on-screen digitisation. A simple vector was used to digitise lineations, meltwater channels and eskers. Polygons were used to digitise hummocky moraine, transverse ridges, ribbed (Rogen) moraine, glaciofluvial sediment accumulations and the break of slope exhibited by subglacial lineations. Ground-truthing was carried out in areas of complexity to verify the mapped glacial landforms. Stoss and lee forms identified from the geomorphological mapping coupled with published erratic pathways (e.g. Goodchild, 1875; Trotter, 1929; Hollingworth, 1931) are used to interpret flow directions throughout the region.

We present local case studies of bedform overprinting in NW England as examples of the major changes in temporal ice sheet flow directions beneath the central sector of the BIIS. Regional maps of subglacial bedforms will be reported in later papers. Our results indicate significant changes in ice flow through time and a highly dynamic ice sheet dispersal centre, but due to a poorly constrained chronology for the region we can only speculate on the periods of time over which flow switching occurred. For this reason we ran a numerical model of ice sheet inception and decay for the British Isles driven by the Greenland ice core temperature record, in order to independently evaluate the potential for highly mobile ice sheet dispersal and basal flow paths.

### 3.4 Glacial geomorphology

We now present glacial geomorphological maps that demonstrate the complexity of ice flow in four locations around the study region. The major flow events in each location are placed in regional context with respect to changing ice sheet dispersal patterns. Each of these events can be placed into a relative age scheme based upon cross-cutting relationships, and these are then reconciled with the independently generated flow phases from our numerical modelling.

#### 3.4.1 *Eastward flowing ice in the Tyne Gap:*

The Tyne Gap (Figs. 3.1 and 3.6a) has been interpreted as a potential eastwards flowing ice stream of the BIIS (Bouledrou *et al.*, 1988; Everest *et al.*, 2005) with a convergence zone generated by ice input from the Lake District and Southern Uplands (Trotter 1929) and a trunk zone demarcated by a smoothed corridor of terrain in which the W-E aligned bedrock structure has been accentuated by glacial streamlining. Geomorphological mapping of the area, coupled with erratic trains and stoss and lee forms indicates that ice flow is predominantly towards the east (Fig. 3.6a) but the visually striking appearance of the bedrock structure gives a false impression of W-E ice flow persistence through the last glacial cycle. Instead, a series of ice flow shifts are apparent in the overprinting of subglacial bedforms (Fig. 3.6b,c), suggesting a more dynamic flow pattern driven by the changing dominance of the competitive ice dispersal centres in the region. In order to differentiate between bedrock and soft subglacial bedform areas geological and superficial deposit maps (DiGMapGb – 625k downloaded from BGS) were draped over the NEXTMap DEM (see Fig. 3.7). Smooth lineations are interpreted as soft-bedded, whilst lineations formed in areas either devoid of surficial deposit, characterised by a high roughness, or which follow or show evidence of geological structure (e.g. Whin Sill: see Fig. 3.7) are interpreted as hard-bedded (Fig. 3.7). Although the hard-bedded lineations are interpreted as glacially-formed, in terms of deriving flow sets and flow phases they can create complications due to the influence of geological structure, and are therefore ignored in this study.

Cross-cutting patterns show that initial flow through the Tyne Gap was from the SW, from the Lake District (pink arrow, Fig. 3.6c). Flow then shifted to an easterly direction (blue arrow, Fig. 3.6c), indicating that ice dispersal was from the Solway Lowlands and that Scottish, Southern Upland ice became more dominant. This increasing dominance of Scottish ice is evident in the final flow phase in the area, specifically down the North Tyne River from the NW, reflecting a northerly and easterly shift in the Southern Uplands ice dispersal centre (orange arrow, Fig. 3.6c). These flow phases help to explain the indistinct boundary between Scottish and Lake District erratics (Trotter, 1929). Ice flow through the Tyne Gap is likely to have occurred throughout all but the final phase drawdown into the Irish Sea.



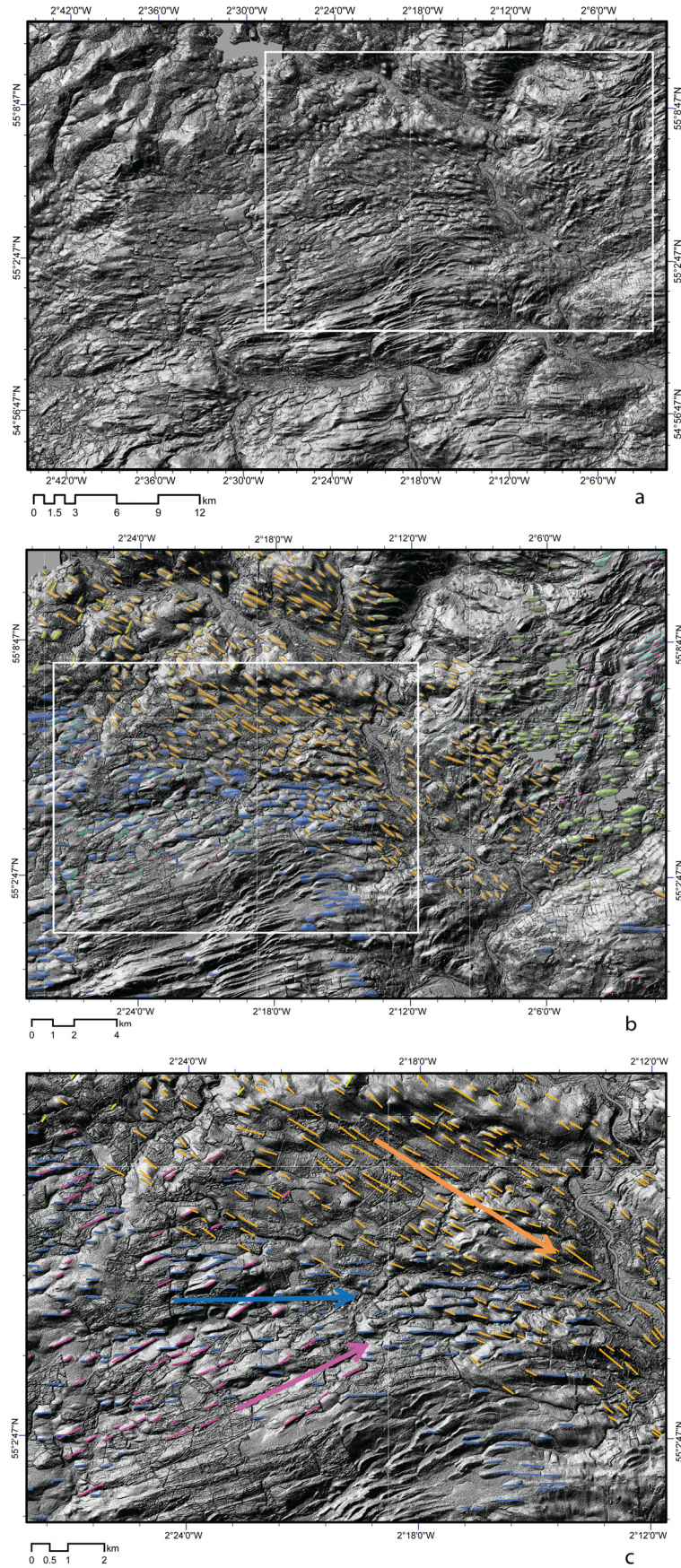


Figure 3.6: Glacial geomorphic evidence of eastward flowing ice in the Tyne Gap (for details see text). Illumination azimuth: 315°.



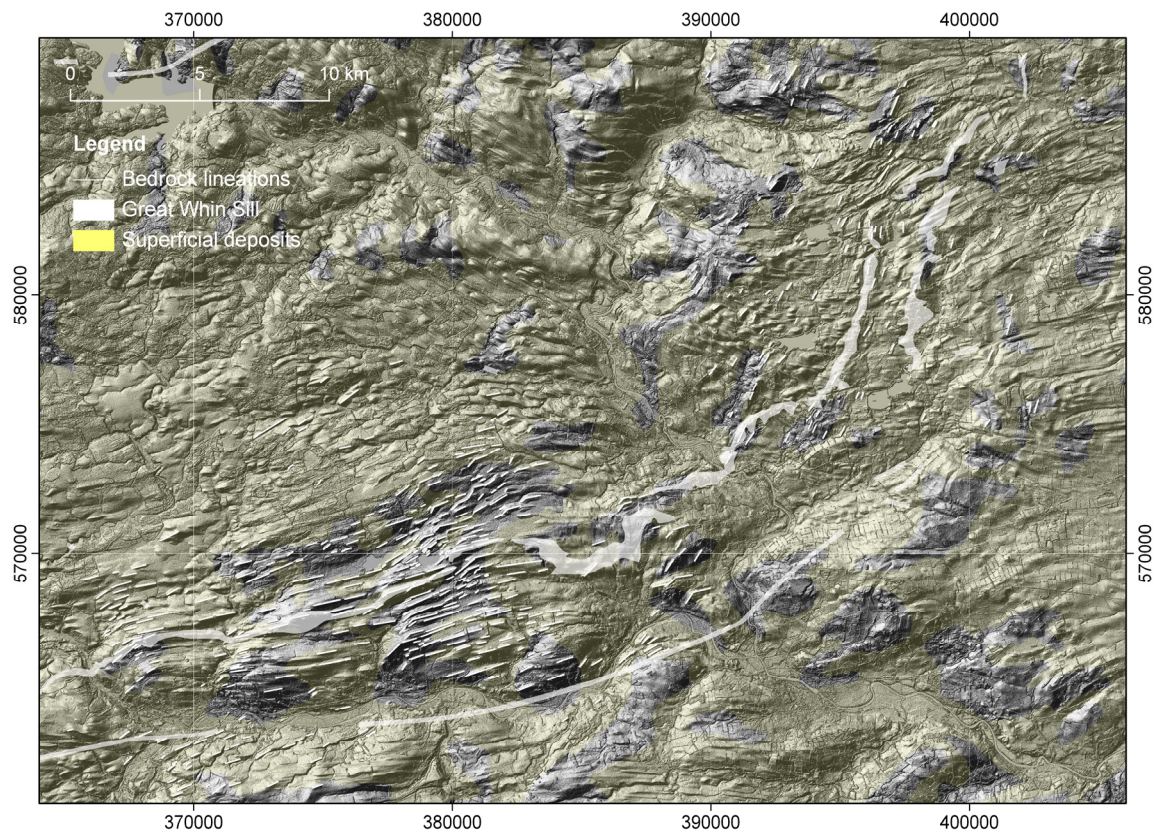


Figure 3.7: Geological influence of bedrock structure on subglacial lineations.

#### 3.4.2 *Opposed ice flow patterns in the Solway Lowlands:*

The Solway Lowlands contains a complex palimpsest of cross-cutting subglacial bedforms (Figs. 3.1 and 3.8) indicative of major shifts in the competing ice dispersal centres of the region. The earliest flow phase is recorded by W-E orientated drumlins, indicative of ice dispersal from a westerly source and generated by the Southern Uplands. This flow direction must have been influenced by the presence of a strong ice dispersal centre of the Lake District and northern Pennines, because the Scottish ice was forced to flow east through the Tyne Gap (blue arrow, Fig. 3.8b,c). The corollary is that an ice divide straddled the outer Solway Firth. A subsequent switch in ice flow is recorded by NE-SW orientated drumlins (yellow arrow, Fig. 3.8b,c), indicating that ice was being drawn down into the Solway Lowlands from southern Scotland. At this time dispersal centres were located over the highland terrain surrounding the Solway Lowlands and the Solway ice divide that was driving ice flow eastwards had dissipated. The final flow phase is recorded by NNE-SSW orientated drumlins, indicative of topographically controlled, deglacial flow from upland sources (white arrow, Fig. 3.8b,c).



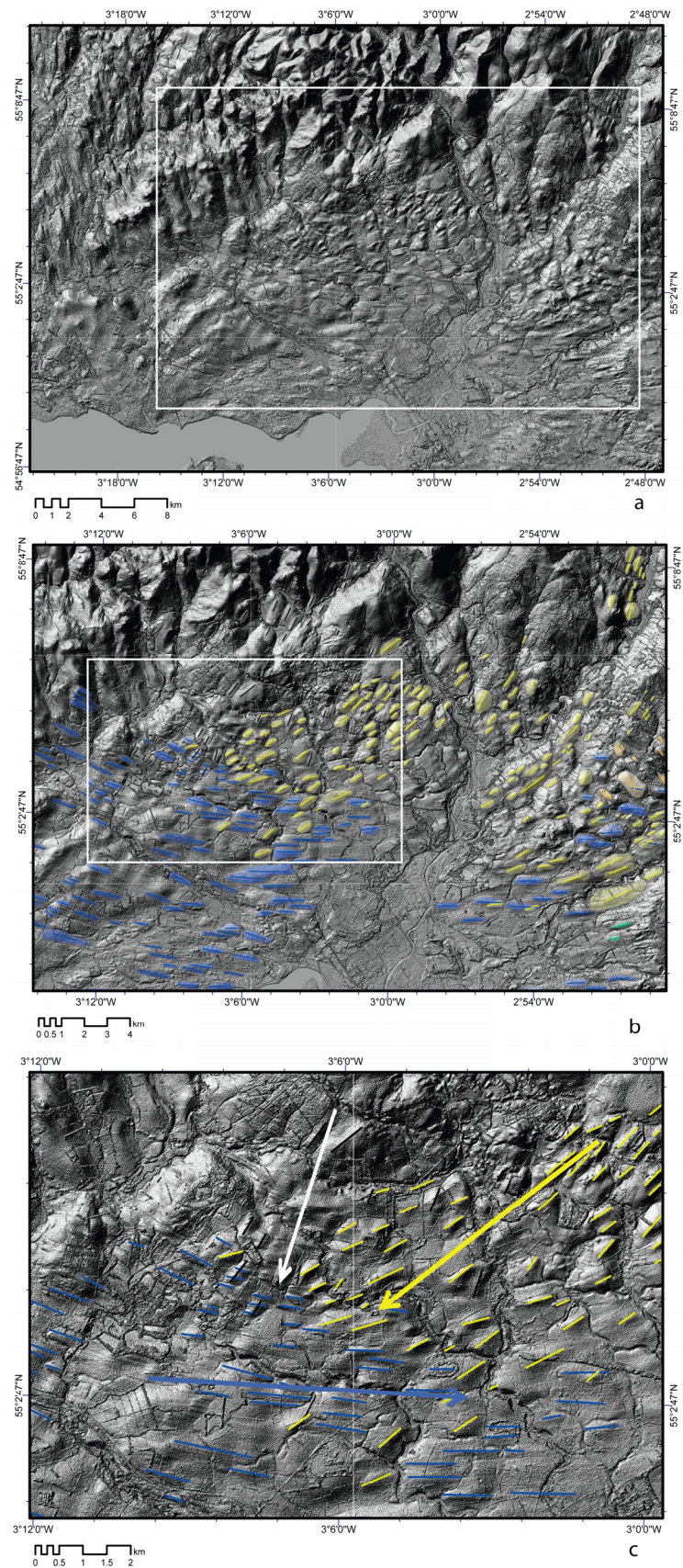


Figure 3.8: Glacial geomorphic evidence of opposed ice flow pattern in the Solway Lowlands (for details see text). Illumination azimuth: 315°.

### 3.4.3 *Bedform overprinting in the Vale of Eden:*

Although much of Edenside is dominated by a SE-NW orientated lineation pattern (Figs. 3.1 and 3.9a), erratic trains (Trotter, 1929; Hollingworth, 1931) and drumlin orientations (Letzer, 1978; Mitchell & Clark, 1994) reveal a complex ice-flow history. In Figure 3.9b, mapping of stoss-and-lee drumlin forms reveals a predominantly north-westerly flow direction (red vectors) in the area to the north and west of Appleby. This is in contrast to drumlin orientations to the area south and east of Appleby, towards the Stainmore Gap. Here the drumlin orientations indicate converging easterly flows (purple, green and blue vectors, Fig. 3.9b) where ice was forced through the Stainmore Gap from the west. These two major flow phases, which have been recognised and grouped based upon the criteria of Clark (1999), are further sub-divided based upon their relative chronology. Superimposed drumlins (Fig. 3.9c) reveal that ice initially crossed Stainmore moving west to east. This was followed by a secondary movement in a north-west direction down Edenside towards the Solway Lowlands. Drumlins formed by the initial easterly flow are preserved in Stainmore but become increasingly rare to the west (Fig. 3.9b,c) where the later flow exerted a greater influence in remoulding the landscape. This switch in ice flow is supported by erratic and till evidence; with Shap granite, Dufton Pike granite and Permian Brockram dispersal trains traced eastwards across Stainmore Gap from the Lake District (Trotter, 1929; Hollingworth, 1931), whilst the northerly flow is constrained by an extensive red till containing Borrowdale lavas from the Lake District (Trotter, 1929). The northwesterly direction of the later stage of ice flow is consistent with inset sequences of glacial deposits (kames; Huddart, 1970) and meltwater erosional channels (Arthurton & Wadge, 1981; Clark *et al.*, 2006; Greenwood *et al.*, 2007) that record a south-easterly recession direction by the last ice to occupy the Eden Valley.

### 3.4.4 *Late stage unconstrained flows in the Solway Firth:*

A very prominent late stage westerly flow of ice down the Solway Firth is documented by well developed drumlins on the lowlands that flank the northern Lake District (Figs. 3.1 and 3.10). The arcuate shape of the drumlin swarm associated with this westerly flow reveals that it was driven by ice flow down the Vale of Eden and driven by an ice divide that connected the Lake District and Pennine dispersal centres. Scottish ice must have been drawn-down in the same direction at this time, because an easterly flow through the Tyne Gap would be glaciologically implausible (i.e. can't have two ice flows moving past each other in opposite directions). The strong westerly lineation pattern has been masked in the west by a large depositional landform in the vicinity of the town Aspatria (Fig. 3.10b, c). This landform comprises a flat-topped ridge which is separated from a heavily pitted terrain by a steep ice-contact NW face (black dashed line, Fig. 3.10c). A quarry in



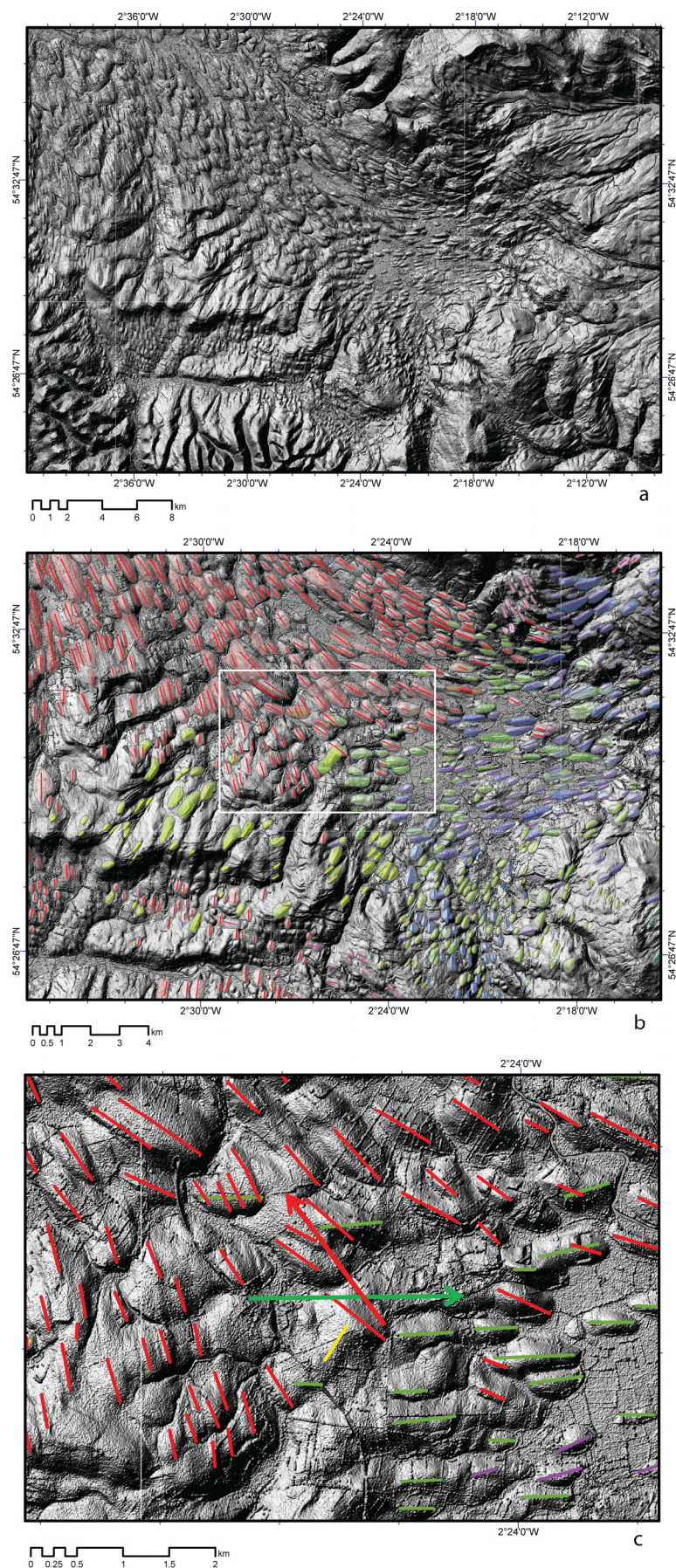


Figure 3.9: Glacial geomorphic evidence of bedform overprinting in the Eden Valley (for details see text).  
Illumination azimuth: 315°.



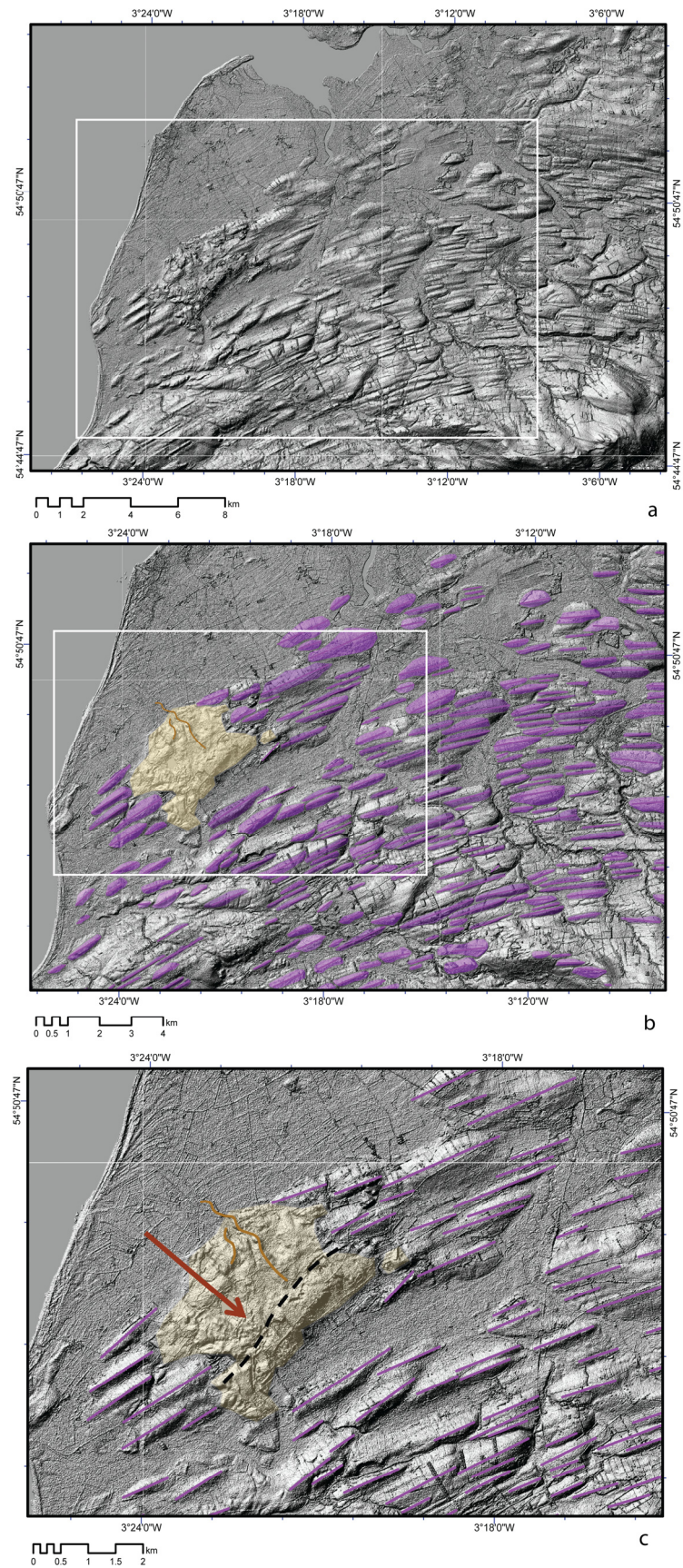


Figure 3.10: Glacial geomorphic evidence of late stage unconstrained flows in the Solway Firth (for details see text). Illumination azimuth: 315°.

the flat-topped ridge displays Gilbert-type foreset beds with palaeo-current indicators that record palaeo-discharges from the NW. The heavily pitted terrain contains NW-SE trending esker ridges indicative of drainage through a former glacier snout towards the ice-contact slope of a delta (Huddart 1970). The delta foresets and pitted terrain are interpreted as the products of ice-marginal deposition into an ice-dammed lake, produced when Scottish, Southern Upland ice re-advanced across the Solway Firth and blocked the local drainage off the northern slopes of the Lake District uplands (Huddart 1970, 1971, 1991). At this time ice flows in the region were driven by dispersal from upland terrains, which resulted in unconstrained advance into the surrounding lowlands. The resulting piedmont lobes were responsible for damming of glacial lakes in the Solway Lowlands, as evidenced by extensive glaciolacustrine sediments at the top of the Quaternary stratigraphic sequences in the region (Huddart, 1971).

#### 3.4.5 *Regional summary of major ice flow phases:*

Based upon the four localised case studies above we identify four major flow phases in the region (Fig. 3.11), acknowledging that further intervening flow phases most likely exist and that not all regional flow phases are represented in every case study. Flow phase I identified predominantly from erratic trains in earlier work (e.g. Trotter, 1929), involved a dominant Scottish dispersal centre, as documented by the transport of Criffel granite erratics to the Eden Valley and the forcing of Lake District ice eastwards over Stainmore. This event is not recorded in the Tyne Gap record (Fig. 3.6) probably because later flows were also aligned W-E but we have indicated easterly flow through Tyne Gap at this time as this would be implicit in the regional flow patterns. Flow phase II involved easterly flow of Lake District and Scottish ice through the Tyne Gap (LT1-3: Chapter 2) and Stainmore Gap (ST1-3: Chapter 2) with an ice divide located over the Solway Firth. This explains the changing ice flow directions in the southern Vale of Eden and over the Tyne Gap. In the latter case, the shift of ice flow from northeasterly to easterly to southeasterly (Fig. 3.6) reflects the dissipation of the Solway Firth ice divide and the re-establishment of Scottish flows across the region, especially with the development of a North Tyne ice flow (LT4: Chapter 2). Flow phase III involves a dominant westerly flow from upland dispersal centres into the Solway Lowlands and along the Solway Firth due to draw-down of ice into the regional topographic low of the Irish Sea Basin (LT5: Chapter 2). This is recorded by strong lineation development in the Vale of Eden (Fig. 3.9), Solway Lowlands (Fig. 3.8) and Solway Firth (Fig. 3.10). Finally, flow phase IV documents the unconstrained late advance of Scottish ice across the Solway Firth, clearly demarcated by the ice-contact delta at Aspatria (SF1: Chapter 2) but also probably recorded by localized valley constrained ice flows to the north of the Solway Lowlands (Fig. 3.8b,c).



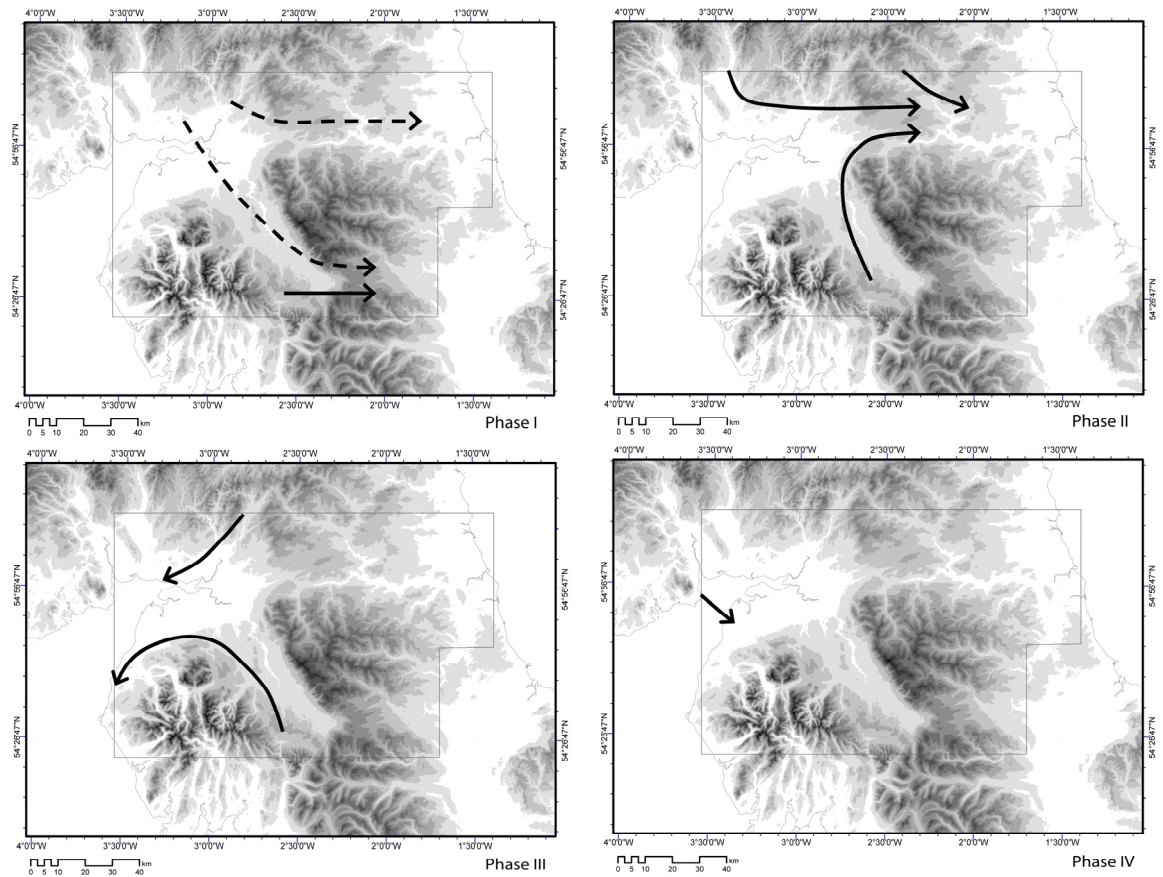


Figure 3.11: Summary maps of major flow events identified in the subglacial bedform mapping. Topography is derived from the SRTM dataset and the box represents the area of mapping.

### 3.5 Ice flow reconstructions from numerical modelling

#### 3.5.1 Model description:

The model we develop here does not aim to reproduce an accurate reconstruction of BIIS history but rather allows us to investigate the effects of ice dynamics on ice sheet evolution and in particular their potential role in changing ice flow directions. Thus, the model includes the minimal requirements for a dynamic ice sheet model, specifically a 2-dimensional ice flow and free surface evolution. By ignoring temperature evolution within and at the base of the ice and any effects of longitudinal stresses our numerical ice sheet model is there for less sophisticated than previous attempts for modelling the BIIS or the Younger Dryas ice cap (Boulton & Hagdorn, 2006; Hubbard, 1999; Gollledge *et al.*, 2008).

A standard time-dependent 2-dimensional shallow ice approximation model (SIA) is used to simulate ice sheet evolution which is based on the 2-dimensional continuity equation for ice thickness. The flux in both horizontal directions is calculated using the SIA and Glen's flow law (Glen, 1955). At the bed, basal motion is either assumed to be zero or related to the basal shear

stress to the power of 3 as similarly used for modelling the Loch Lomond Re-advance by Hubbard (1999). A constant rate factor is used for the ice (isothermal) of  $3.2 \cdot 10^{-24} \text{ Pa}^{-3} \text{ s}^{-1}$  corresponding to an ice temperature of  $-2^\circ\text{C}$  (Paterson & Budd, 1982) and an additional enhancement factor of 4 is applied to further soften the ice, as often used in ice sheet modelling (Ritz *et al.*, 2001; Boulton & Hagdorn, 2006).

We account for the expected higher basal motion resulting from enhanced deformation of ocean sediments by enhancing the sliding coefficient in our sliding relation by a factor of 10 (from  $6.4 \cdot 10^{-14} \text{ Pa}^{-2} \text{ a}^{-1} \text{ m}$  to  $6.4 \cdot 10^{-13} \text{ Pa}^{-2} \text{ a}^{-1} \text{ m}$ ) in areas which are below sea level at present. Thus, for the same basal shear stress, the resulting sliding velocity over ocean areas is 10 times higher than over land-based areas. A simple local relaxation scheme is used for isostatic bedrock adjustment (Oerlemans, 1980). At the marine boundary of the ice sheet a flotation criterion has been applied that basically removes any ice below flotation (van der Veen, 1996; Vieli *et al.*, 2001).

The mass balance is calculated from considering the annual balance between accumulation and ablation. Ablation is calculated using a positive degree day model (Reeh, 1989) with a correction of temperature relative to the present day temperature (average from 1971-2000) to account for elevation changes of the ice sheet surface. Furthermore, temperature perturbations can be applied with time for forcing changes in mass balance.

Accumulation is estimated using the present day precipitation field (average from 1971-2000) as a reference and applying several corrections. Altitude and the prevailing wind direction from the west strongly affect the present precipitation pattern of the British Isles. These effects have been accounted by a correction for altitude change relative to the present surface elevation and a correction for the changing surface slope in the prevailing wind-direction, as used by Payne & Sugden (1990) for moisture calculations. A further correction is applied to account for the temperature dependency of precipitation as a result of reduced moisture content in the air with decreasing temperature (Marshall *et al.* 2002; Huybrechts *et al.*, 1991). Overall, our model for mass balance involves 6 model parameters which have been chosen within the suggested values from the literature and if necessary have been further constrained by the basic tuning process below.

### 3.5.2 *Settings and tuning of model parameters:*

The horizontal grid-size is 10 km and a time-step of 10 years has been used. The mean surface topography over each 10 km x 10 km grid cell has been taken as input surface topography and has been corrected by adding one standard deviation of each grid cell to correct for the smoothing effect in mountainous areas from the gridding processes. The prevailing wind-direction is assumed to be from 250 degrees (approximately WSW) and is kept constant over time. An implicit finite-difference scheme is used to predict the thickness evolution. The forcing in temperature anomaly

(Fig. 3.12a) is derived from the  $\delta^{18}\text{O}$  GRIP ice-core as provided for the EISMINT experiments (Dansgaard *et al.*, 1993; Huybrechts, 1998) and is used as input to determine mass balance. Sea level has been kept constant at current level over the whole model run period.

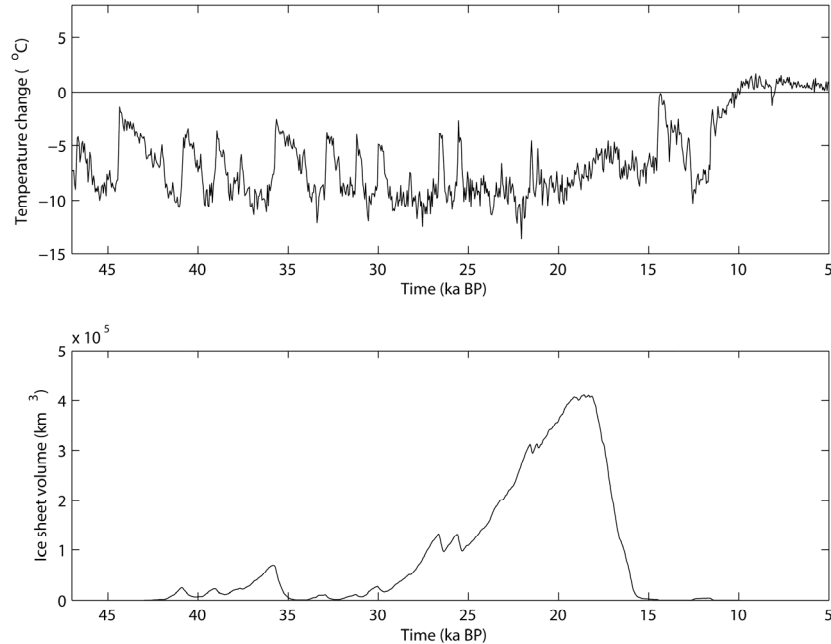


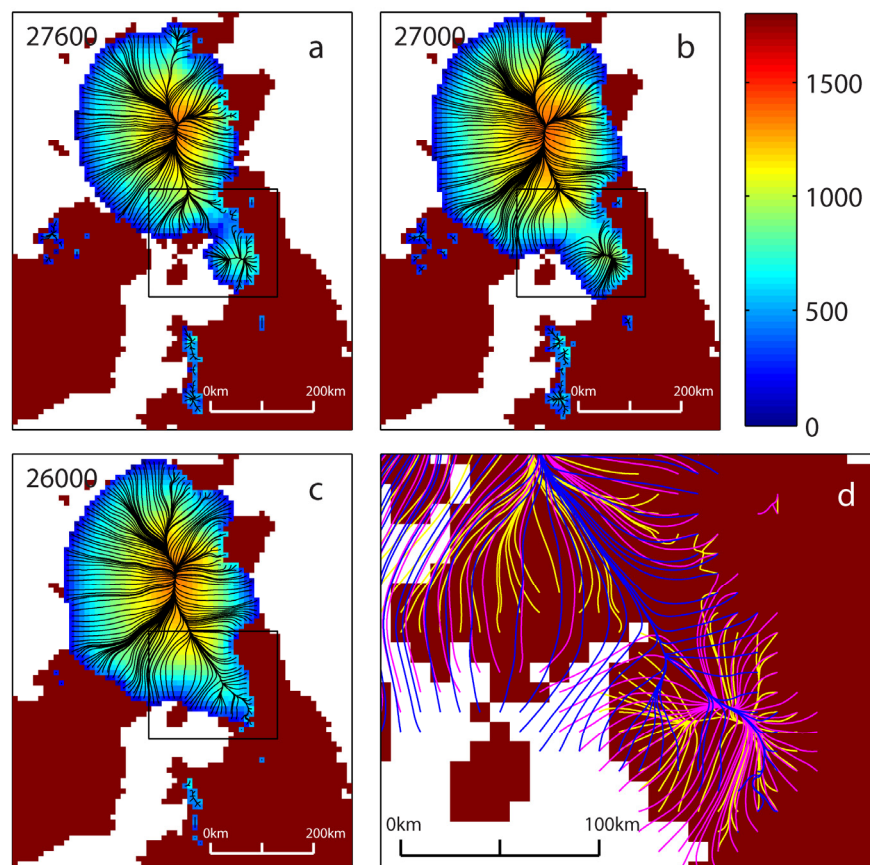
Figure 3.12: (a) Temperature anomaly from present with time derived from  $\delta^{18}\text{O}$  ice core record (Dansgaard *et al.* 1993; Huybrechts 1997) which is used as forcing the mass balance input for the ice sheet model; (b) modelled ice sheet volume of the BIIS over the last 45 ka.

Using the GRIP ice core temperature-forcing, the model has been run over the last 43 ky BP. The model parameters within the mass balance model, the rate factor and the sliding coefficient have then been tuned to match three main constraints: First, to match the extent of the Younger Dryas ice coverage; second, ice free conditions before the Younger Dryas event; and third an approximate agreement with the current knowledge of ice sheet extent (cf. Bowen *et al.*, 2002; Clark *et al.*, 2004; Evans *et al.*, 2005; Ó Cofaigh & Evans 2007). This provided good constraints for most of the parameters within the model. We deliberately did not further tune the model to match specific evidence from geomorphology such as our observed flow direction sets (Fig. 3.6-3.11), because our aim was to independently investigate the effect of ice dynamics on flow directions rather than to reproduce the observations. The results are also in general agreement with earlier modelling reconstructions for the LGM (Boulton & Hagdorn, 2006) and the Younger Dryas ice sheet (Hubbard, 1999; Payne & Sugden, 1990). An additional model run starting from 120 ky BP also showed that the start point of 43 ky is justified as it turned out to be ice free at that time. Also, the

same model run with lower horizontal grid-resolution using 20 km showed qualitatively very similar results.

### 3.5.3 *Results and discussion of modelling:*

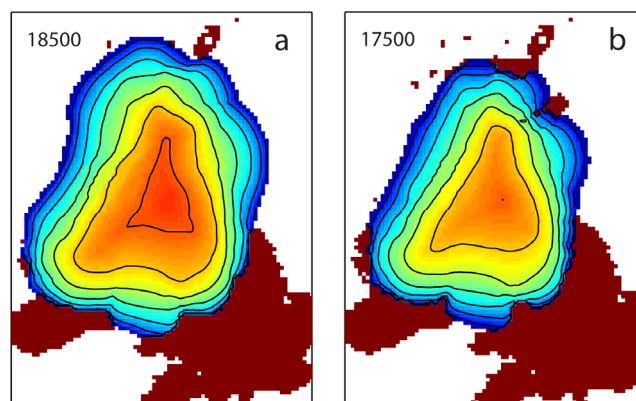
The modelled history of the BIIS is summarized in terms of ice volume in Figure 3.12a and shows fluctuations of smaller ice caps before 30 ka BP, followed by a more or less continuous growth into an extensive LGM ice sheet. The forcing mechanism for this behaviour appears to be the length of the period of significantly lower temperatures; the model requires a long enough period of significantly lower temperatures to grow to its full LGM extent (Fig. 3.12). Most importantly with respect to NW England and SW Scotland, the model grows initially upland ice masses over the Lake District/Pennine area from which ice flows radially and then around 27 ka BP joins the Scottish ice cap (Fig. 3.13). The extensive period of continuously low temperatures from 26 to 19 ka BP then allows the ice sheet grow to its full LGM extent (Fig. 3.12 and 3.14a). The LGM-extent is reached at 19.5 ky BP and is maintained for only 1000 years, with recession taking place after 18.5 ky BP and then proceeding very rapidly over the next 2500 years (Fig. 3.12 and 3.14). The recession from the LGM also displays a complex pattern (Fig. 3.15), with distinctly faster retreat rates in areas currently occupied by ocean and the development of several individual upland ice domes towards the end of deglaciation.

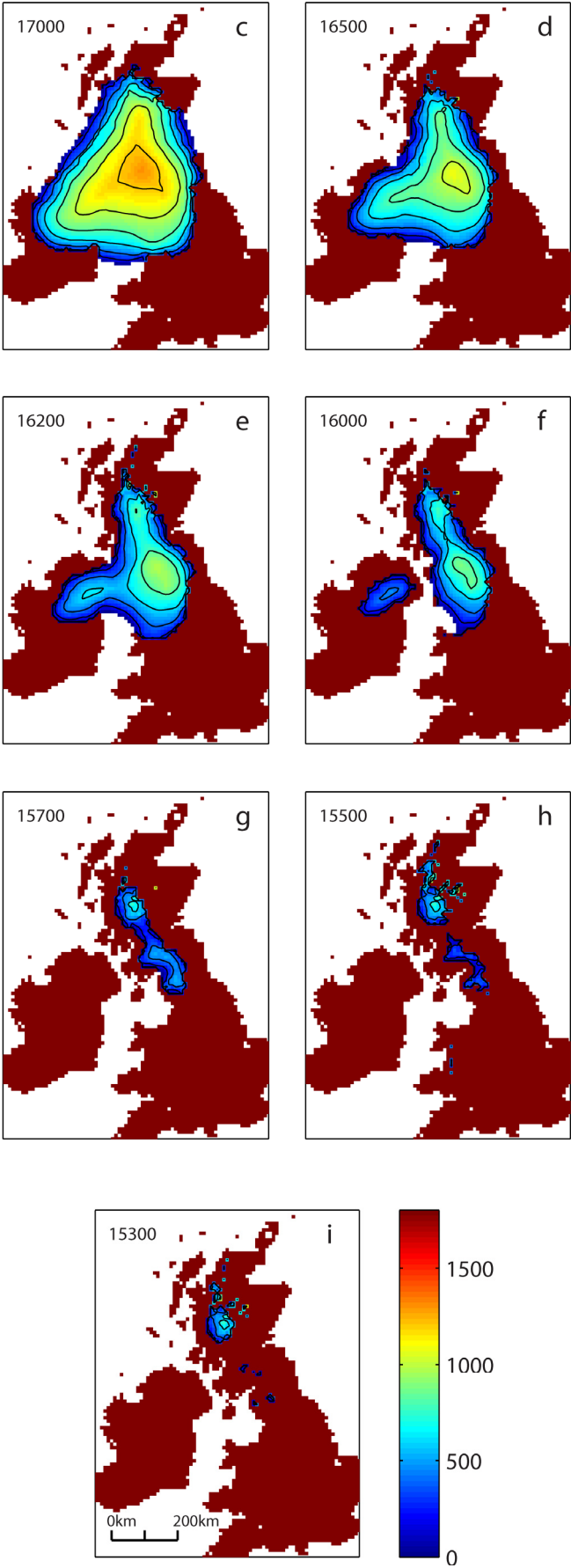


< Figure 3.13: (a-d) Maps of modelled surface elevation (colours in m a.s.l.) and ice flow trajectories (black lines) of the BIIS during build up to the LGM, for the labelled calendar years before present, through the coalescence of upland icefields from 27.6 to 26 ka BP. The black rectangular frame indicates the outline of the zoomed area in Figure (d) which shows a map of overlain flow trajectories from Figure (a-c): yellow lines = 27.6 ka BP, pink lines = 27.0 ka BP, blue lines = 26.0 ka BP. The brown areas indicate the present land surface.

The modelled flow trajectories at different time slices during the recession from the LGM reveal significant ice flow directional switches over NW England/SW Scotland (Figs. 3.15 and 3.16). Particularly significant are the relatively short timescales (few hundreds of years) over which these shifts in flow pattern occur. The modelled retreat pattern and the ice-flow trajectory changes seem to be a result of dynamical processes: firstly, the prescribed enhanced ocean area leads to a more efficient mass transfer and therefore preferential draw down of the ice surface over such areas; secondly, calving leads to accelerated retreat over marine over-deepenings; and thirdly, and important for the ice flow directions towards the end of the deglaciation, is the effect of the underlying bed topography once the ice sheet has thinned substantially. We have validated this positive relationship between ice flow acceleration and marine re-entrants by running additional models with the same, land-based sliding coefficient over land and ocean surfaces, and these have revealed far less variable retreat rates and less pronounced flow direction changes over time.

Due to the coarse resolution of the grid cells in our numerical ice sheet model, it is difficult to make firm conclusions on localized ice flow patterns in the region. However, changes in regional flow pathways are clearly reproduced by the model, highlighting the switches in basal ice flow trajectories that are associated with mobile ice divides. This mobility and flow switching allows us to have confidence in our reconstructions of rapid temporal overprinting of the subglacial bedforms of NW England and SW Scotland (cf. Salt & Evans, 2002).





< Figure 3.14: (a-i) Maps of modelled ice surface elevation (contours and colours in m a.s.l., 200m contour interval) of the BIIS during retreat from the LGM for the labelled years before present and postdating 18.5 ka BP. The brown areas indicate the present land surface.

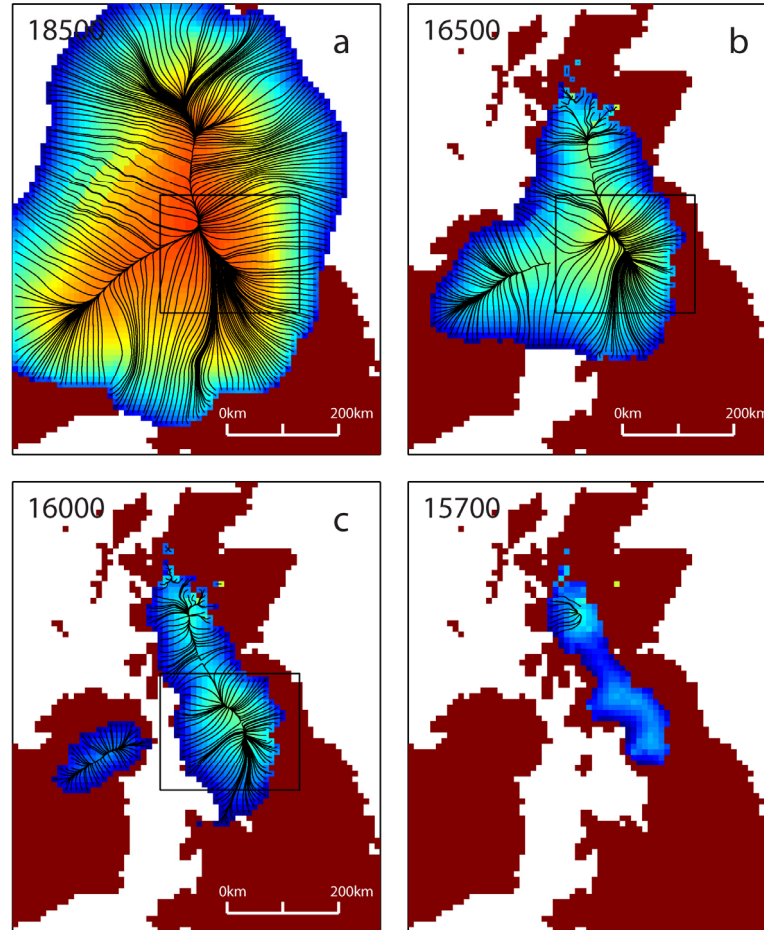


Figure 3.15: (a-d) Maps of modelled surface elevation (colours in m a.s.l.) and ice flow trajectories (black lines) during deglaciation of the BIIS for the labelled years before present. The black rectangular frame indicates the outline of the zoomed area in Figure 3.16.

Some chronological control based on radiocarbon dated till stratigraphies (expressed from here on in calibrated years) around the Irish Sea Basin, the area into which the NW sector of the BIIS sheet was draining, allow us to independently reconcile the developmental stages of our sheet model. The southernmost limit ice sheet advance has been dated on the SE coast of Ireland by Ó Cofaigh and Evans (2007) at < 23.9 ka, when a large ice stream drained southwards down the Irish Sea Basin from the centre of the BIIS (Evans & Ó Cofaigh, 2003; Boulton & Hagdorn, 2006). By the time of the Clogher Head re-advance at 18.3-17.0 ka and the Killard Point Stadial some time after 17.0 ka (McCabe & Clark, 1998, 2003; McCabe *et al.*, 1998, 2005, 2007), the ice sheet margin had receded



to the northern end of the Irish Sea Basin and was fed by outlet lobes draining through the major re-entrants such as the Solway Lowlands. At this time ice would have been streaming westwards from the upland dispersal centres in response to draw down into the deepening waters of the Irish Sea (cf. Eyles & McCabe, 1989). Our numerical model depicts south-south westerly ice flow to ice margins lying over the Irish Sea Basin during the 18.5 and 16.5 ka time slices (Figs. 3.14 and 3.15), the latter being located in the northern part of the basin in the vicinity of the Killard Point Stadial ice-marginal landforms and sediments. We therefore have confidence that our model, bearing in mind that it has a coarse spatial grid resolution, is a reasonable fit to the reconstructed late Quaternary history of the region. This allows us to make some realistic conclusions about rates of ice flow trajectory changes during the later stages of the last glacial cycle.

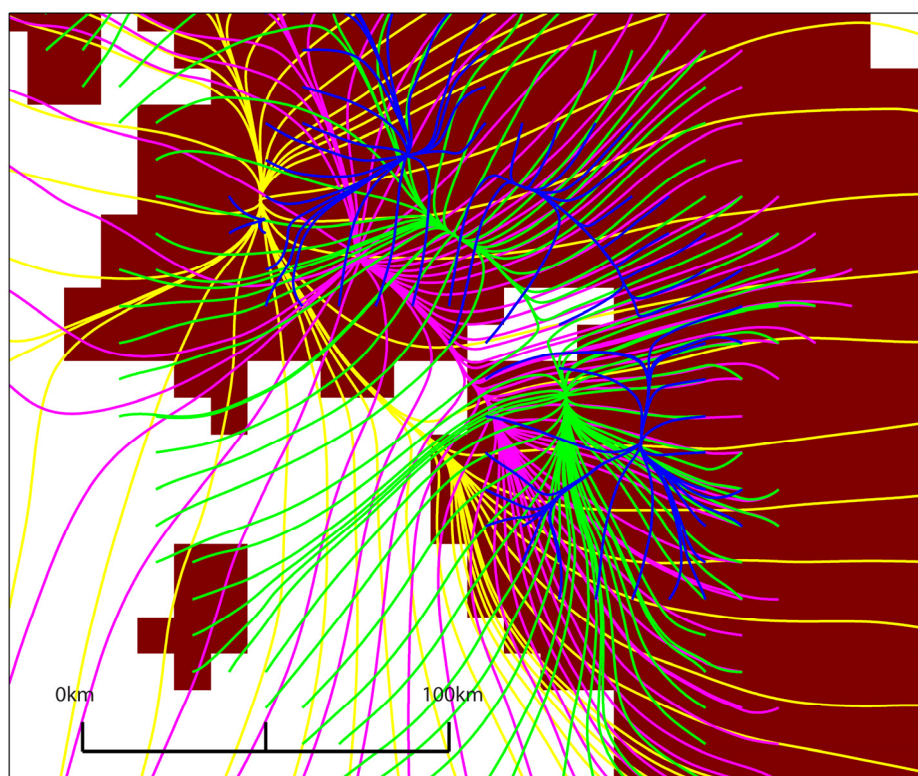


Figure 3.16: Map of overlain ice flow trajectories of the 4 flowsets for the different times slices in Figure 3.15: yellow lines = 18.5 ky BP, pink lines = 16.5 ky BP, green lines = 16.0 ky BP and blue lines = 15.7 ky BP.

### 3.6 Integration of modelling and glacial geomorphology

Both palaeoglaciological reconstructions (e.g. Dyke & Morris, 1988; Boulton & Clark 1990a, b; Clark, 1993, 1997; Clark & Meehan, 2001) and modern ice sheet observations (e.g. Bamber *et al.*, 2000) have clearly demonstrated that basal flow can be complex and subject to significant switching over short timescales. This has been demonstrated also for the central part of the BIIS by Mitchell

(1994, 2007), Salt and Evans (2002) and Mitchell and Riley (2006) and so it is unsurprising that the subglacial bedform record of NW England reflects the dynamic and mobile nature of the ice sheet during the last glacial cycle. This reflects the tendency for ice sheets to preserve geomorphological flow features (Kleman, 1994), revealing a palimpsest of cross-cutting relationships (Boulton & Clark 1990a, b; Clark, 1993) and therefore, a series of flow phase relationships.

The ability to preserve landforms is pervasive within the ice-sheet system (Kleman, 1994) and occurs when basal shear stress is lower than bed strength (Clark, 1999). This can relate to cold-based ice (Kleman & Borgstrom, 1994), low velocity such as at ice divides, and shallowing of the deformation layer (Clark, 1999), as well as the period of time that the sediment is exposed to streamlining (Kleman & Borgstrom, 1996). Kleman (1994) also infers that preservation of more robust landforms like drumlins can be ubiquitous with warm based conditions. In north-west Sweden Kleman (1992) was able to reconstruct a relative chronology and basal thermal regime history of flow events using morphological and cross-cutting criteria. Formation of composite subglacial landform assemblages has been recognised for a long time (Rose & Letzer, 1977) and it is now recognised that cross-cutting refers to an intermediate stage in the complete re-organisation of the bed (Clark 1994). The frequently pervasive occurrence of such cross-cutting suggests that the erosive nature of ice sheets has been overestimated in the past, and ice in fact mainly re-distributes sediment to various levels of attenuation dependent on such factors as substrate rheology, basal thermal regime and duration of flow phases. However, deciphering the numerous and often rapid changes in ice flow directions during even one glacial cycle, due to the spatial and temporal variability in subglacial bedform modification, requires us to make difficult correlations between localised swarms of overprinted features. The large areas of lowland topography in NW England have imparted warm-based, sliding characteristics on large areas of the BIIS bed in the region, giving rise to a complex overprinted bedform signature from which we have deciphered regional flow complexities (Fig. 3.11). We now attempt to reconcile our ice flow phases with broader scale ice flow histories and our independently constructed numerical model. Given that bedform preservation will deteriorate with age, especially as similar ice flow trajectories have affected most areas in the study region more than once through the last glacial cycle, we have worked chronologically backwards through the numerical model when assigning relative ages to the bedform flow phases. In summary, all of our flow phases can be accommodated by the post 18.5 ka BP period in the numerical model, although there is a possibility that some bedforms might have survived from previous flow phases.

Our numerical ice sheet model indicates that the BIIS developed an elongate, triangular-shaped dome at the LGM, centred over SW Scotland and widening towards the northern part of the Irish Sea Basin and NW England and coalescing with the Irish sector of the ice sheet. The longest, N-S orientated ice divide on this dome migrated eastwards between 18.5 and 16.5 ka to be located over the highlands of NW England and SW Scotland, where it was connected to residual Irish ice by a

saddle over the north Irish Sea Basin (Fig. 3.14). Our modeled time slices of 18.5, 16.5, 16.0 and 15.7 ka (Fig. 3.15 and 3.16) allow us to assess the impact of this ice divide shift on regional flow trajectories, although the coarse resolution of the underlying topographic grid restricts us from making firm correlations between model time slices and geomorphologically defined flow phases.

At 18.5 ka, the westward location of the ice divide resulted in the driving of ice flow towards the east across the study region. This resulted in the incursion of Scottish ice into the northern Pennines and vigorous flow along the Tyne Gap. The easterly orientated bedforms of flow phases I through II (Fig. 3.11) were most likely moulded at this time, although the phase I southerly incursion of Scottish ice to the southern end of the Vale of Eden can not be accommodated in the modelled ice configuration, because Lake District ice is involved in the vigorous easterly flow. Rather, it is the later time slice at 16.5 ka when the Lake District part of the dispersal centre initiates some radial flow over NW England and produces some southeasterly ice flow up the Vale of Eden. Although this could have potentially transported Scottish erratics southwards, it is most likely that the Scottish material had been delivered previously to the Solway Lowlands from where it was then carried by more local ice; flow of Scottish ice up the Vale of Eden and through Stainmore is glaciologically implausible in our numerical model. Although we have no ages for the Scottish erratics in the Vale of Eden, they could have been transported initially to NW England during a previous glaciation (glacial lake sediments between Scottish and Eden Valley tills at Langwathby do record a break in ice coverage at some time) and/or during Salt and Evans's (2004) southerly and southeasterly flows. Even though some of Salt and Evans's (2004) flow stages may not date to the last glacial cycle, they can still be accommodated in our numerical ice sheet model. Specifically, the southeasterly flows on the east side of the Southern Uplands may relate to the 18.5 ka time slice when the ice divide was driving regional Scottish ice in a southeasterly direction over the region; this was also a flow direction forced upon the Southern Uplands ice mass by the more powerful Highland ice flowing southwards from the Firth of Clyde (Salt & Evans, 2004). This vigorous southerly flow of Highland ice is recorded on the Isle of Man, where Roberts *et al.* (2007) identify an early flow phase when Scottish Highland ice moved in a southeasterly direction, pinning Southern Uplands and Lake District ice masses against the Cumbrian coast. The later easterly migration of the Southern Uplands ice divide clearly initiated southerly to southwesterly flow over NW England until 16.0 ka, explaining the development of subglacial bedforms during Stage A towards the southwest, pre-Stage B towards the south, Stage B towards the southwest and Stage C towards the south-southwest. This change in flow dominance is recorded also on the Isle of Man where Roberts *et al.* (2007) report a younger phase of south-westerly orientated striae, erratic trains and bedforms; the change in ice flow direction over the Isle of Man is accommodated in the 18.5 and 16.5 ka time slices from our numerical model.

The flow of both Lake District and Scottish ice along the Tyne Gap during phase II is best replicated by the 16.5 and 16.0 ka time slices. The likely changes in ice flow directions through phase II are

recorded on the north side of the Tyne Gap where the shift of ice flow from northeasterly to easterly to southeasterly reflects changing dominance between Lake District and Scottish ice input to the regional easterly flow and also the dissipation of the Solway Firth ice divide and re-establishment of Scottish flows across the region. Our model, however, does not reproduce this particular flow shift, which may be due to the spatial and temporal resolution of the event. At 16.0 ka the draw-down into the Solway Firth resulted in very vigorous ice streaming over the northern coastal fringes of the Lake District, giving rise to the production of the strongly drumlinised terrain associated with flow phase III. This flow phase is now thought to represent a re-advance of ice following the development of an ice free enclave and pro-glacial lake formation in the Solway Lowlands (Livingstone *et al.*, sub) which is not reproduced by the model.

The 15.7 ka time slice shows not only an easterly migration of the main ice divide to the east of the Solway Lowlands but also the establishment of localised upland dispersal centres from which ice flowed radially down the Vale of Eden and the Southern Upland valleys into the fringes of the Solway Lowlands (Stages F-G of Salt & Evans, 2004). Ice margins at this time are recorded by significant ice-marginal landform-sediment assemblages in glaciated valley settings, such as those demarcating the Eden Valley glacier at the “Brampton kame belt” and associated Pennine escarpment meltwater channels (Arthurton & Wadge, 1981; Clark *et al.*, 2006; Greenwood *et al.*, 2007). Overprinted drumlins have been used to verify the late reversal of ice flow in the Vale of Eden (Mitchell & Riley, 2006) and to reconstruct the local ice divide migration over the western Pennines and Howgill Fells (Mitchell, 1994). Either at this time or earlier (i.e. immediately after flow phase III), an unconstrained re-advance of Scottish ice impinged upon the southern shore of the Solway Firth to construct the Aspatria ice-contact delta. Although we have assigned this flow phase IV, our numerical model does not replicate any such re-advance, making it likely to be a late stage dynamic ice marginal response to rising sea levels and/or early dissipation of Lake District ice flow in the Firth. In the circumstances a surge origin can not be dismissed, explaining why the climatically driven model does not replicate phase IV. Moreover, Salt and Evans (2004) have previously identified a late stage re-advance by Scottish ice into Loch Ryan (Stage G, “Stranraer Re-advance”), indicating that pulsed recession/surging may have been a characteristic of the Highland ice as it retreated in contact with deepening marine water along the Scottish coast.

Our knowledge of the patterns of glacierization of landscapes with diverse topography allows us to make some realistic interpretations of the early stages of ice flow in the study region, which can in turn be reconciled with both existing data on erratic dispersal and our numerical model. The well established principle of “instantaneous glacierization” (Ives *et al.* 1975) implies that upland landscapes such as the Lake District fells, the North Pennine plateaux and the Southern Uplands will have developed an ice cover first, leading to the radial flow of valley glaciers into the lowlands of the Vale of Eden, Tyne Gap and Solway Firth. This ice would have effectively drained unimpeded into the northern part of the Irish Sea Basin until Scottish Highland ice advanced far enough south to

wrap around the Southern Uplands ice mass and up against the Cumbrian west coast, causing NW English/SW Scottish ice over the Solway to thicken as illustrated by the numerical modelling in Figure 3.13. This thickening would have caused regional ice flow to reverse, invigorating easterly flow along the Tyne Gap and forcing ice up the Vale of Eden and over Stainmore, thereby wrapping around the plateau-based local ice of the Cross Fell area and forcing it to flow east (Mitchell 2007). During deglaciation this sequence of flow switching should logically be reversed, although some re-advances may have resulted in localised flow adjustments as glacier margins were drawn down into proglacial lakes or the deepening sea. An example of the reversal of the ice sheet inception pattern is the final phase of deglaciation in the Vale of Eden, which was characterised by ice recession southwards (our phase III) and onto the North Pennine/Cross Fell plateau, as documented by the inset lateral meltwater channels and kame terraces of the Melmerby-Brampton area (Arthurton & Wadge, 1981; Clark *et al.*, 2006; Greenwood *et al.*, 2007).

### 3.7 Conclusions

A number of important conclusions have arisen from our dual approach to deciphering the palaeoglaciology of the central sector of the BIIS. Firstly, with respect to the glacial geomorphology of the region:

- Systematic mapping of subglacial bedforms from four case studies around NW England has identified overprinted subglacial bedforms, which relate to temporally superimposed flow sets indicative of complex ice sheet flow dynamics. The bedform signature of former ice sheet flow is of a complexity similar to that previously reported for the Southern Uplands (Salt & Evans 2004), immediately to the north of the study region but, more significantly, records reversals in glacier flow during glaciation.
- Large scale cross-cutting bedform patterns can only have been produced by significant shifts in ice dispersal centres through time and we have identified four major phases of regional ice flow in order to simply accommodate all the localized overprinting. Flow phase I involved dominant Scottish ice flow and the forcing of Lake District ice eastwards over Stainmore. Flow phase II involved more rigorous flow from the Lake District, Howgill and Pennine uplands, resulting in easterly flow of Vale of Eden and Scottish ice through the Tyne Gap and Stainmore Gap with an ice divide located over the Solway Firth. Flow phase III involved a dominant westerly flow from all upland dispersal centres into the Solway Lowlands and along the Solway Firth and this was locally masked by the deposition of an ice-marginal delta during flow phase IV when Scottish ice re-advanced southeastwards across the Solway Firth.

Secondly, with respect to numerical ice sheet modelling we identified a number of significant factors that complement our geomorphologically-based assessments of palaeo-ice dynamics:

- Modelled ice sheet recession rates are very rapid, with the SW Scotland/NW England sector of the BIIS remaining as a major dispersal centre for only around 2,500 years after the LGM. Our model depicts a dynamic ice sheet with no real steady state and constantly migrating dispersal centres and ice divides. This dynamism is consistent with the ice-flow phases recorded in the superimposed bedform record, and moreover implies that the subglacial streamlining of flow sets was most likely imprinted on the landscape over short phases of fast flow activity and during the recession from LGM, with some flow reversals taking place in less than 300 years.
- The modelled pattern of ice sheet recession reveals greater thinning over oceans and flat lying terrestrial areas and this in turn initiates rapidly evolving flow changes, particularly draw-down into marine/estuarine re-entrants during overall ice sheet wastage.
- Our modelling has produced complex recession and flow change patterns with a very unsophisticated numerical ice sheet model and parameter setting and is governed entirely by internal flow dynamics. It has not been necessary to incorporate “on and off” switches for ice streaming nor have we needed to invoke a complex temperature evolution. It is as such a far simpler model than that of Boulton and Hagdorn (2006) but it still depicts the basic dynamic features necessary to explain the complex and enigmatic subglacial bedform signature of NW England.
- During the short and vigorous deglaciation depicted by our model, there is a strong potential to generate large volumes of meltwater, which in turn feedback into vigorous concurrent subglacial bedform modification and contribute to the carving of dense networks of glaciofluvial meltwater drainage channels, such as those in the Vale of Eden.

### **3.8 Acknowledgements**

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## Chapter 4

# **A major ice drainage pathway of the last British-Irish Ice Sheet: the Tyne Gap, northern England**

Livingstone, S.J., Ó Cofaigh, C. & Evans, D.J.A. In press. A major ice drainage pathway of the last British-Irish Ice Sheet: the Tyne Gap, northern England. *Journal of Quaternary Science*.

## **Abstract**

The Tyne Gap is a wide pass, situated between the Scottish Southern Uplands and the English Pennines that connects western and eastern England. It was a major ice-flow drainage pathway within the central sector of the last British-Irish Ice Sheet. This study presents new glacial geomorphological and sedimentological data from the Tyne Gap region that has allowed detailed reconstructions of the palaeo-ice flow dynamics throughout the Main Late Devensian (MIS 2). Mapped lineations reveal a complex palimpsest pattern which shows that ice flow was subject to multiple switches in direction. These have been summarised into three major ice flow phases. Stage I was characterised by convergent Lake District and Scottish ice that flowed east through the Tyne Gap, as a topographically controlled ice stream. This ice stream was identified from glacial geomorphological evidence in the form of convergent bedforms, streamlined subglacial bedforms and evidence for deformable bed conditions; Stage II involved the northerly migration of the Solway Firth ice divide back into the Southern Uplands, causing the easterly flow of ice to be weakened, and resulting in south-easterly flow of ice down the North Tyne Valley; and Stage III was characterised by progressive westwards retreat westwards back across the watershed, prior to ice stagnation and the strong drawdown of ice into the Irish Sea Ice Basin during the ‘Blackhall Wood re-advance’.

## **4.1 Introduction**

During the Main Late Devensian (Dimlington Stadial) glaciation the central sector of the last British-Irish Ice Sheet (BIIS) was characterised by a complex sequence of flow events as recorded by subglacial bedforms (Livingstone *et al.*, 2008; Evans *et al.*, 2009). This is consistent with recent palaeoglaciological research suggesting that ice-sheet flow through a glacial cycle is highly dynamic (e.g. Dyke and Morris, 1988; Boulton and Clark 1990a, b; Clark, 1997, Clark and Meehan, 2001), with advance and decay characterised by multiphase flow at any one location. Palimpsest landscapes have revealed preservation of successive flow phases (Dyke and Morris, 1988; Kleman, 1994), with subsequent superimposed and cross-cutting bedforms. Despite recent

attempts to rectify a paucity of data on subglacial bedform palimpsests in Britain (Salt, 2001; Salt and Evans, 2004, Clark *et al.*, 2004; Evans *et al.*, 2005, 2009), large regions remain under-researched.

The Solway Lowlands are situated within the central region of the former BIIS (Fig. 4.1) and contain a record of complex, multi-sourced, competing ice flows from the Southern Uplands, Pennines, Lake District and the Irish Sea Basin (Trotter, 1929; Hollingworth, 1931). This inter-regional complexity has implications for: (a) ice sheet configuration, stability and dynamics; and (b) major ice flow arteries through the Tyne Gap, Stainmore Gap and Solway Firth, which acted as route-ways for the transfer of ice to large parts of the margin of the BIIS.

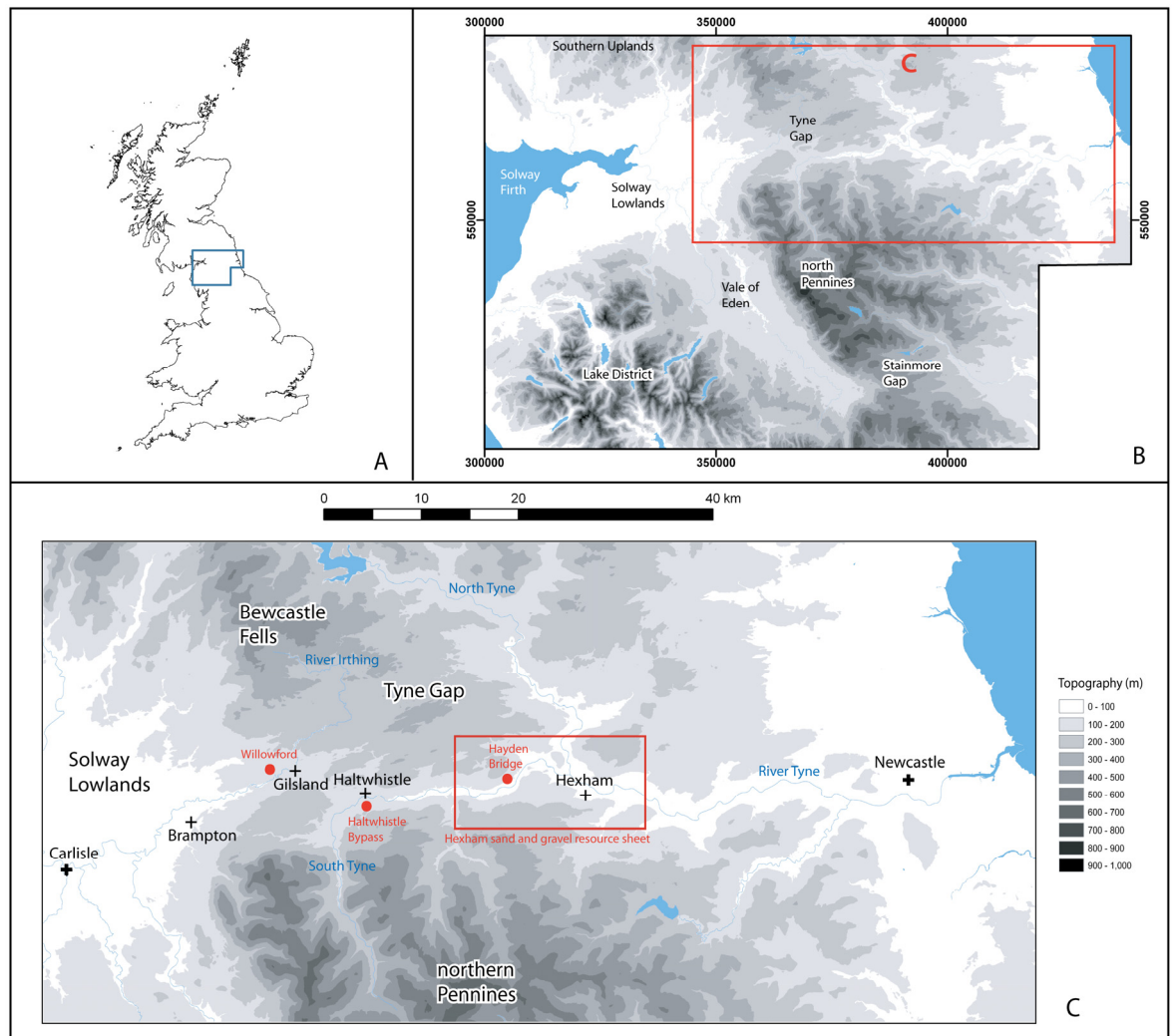


Figure 4.1: (a) Location map of the study area, (b) Topography of the Tyne Gap and the surrounding area, (c) The Tyne Gap: topography, place names, major rivers and field locations (in red).

The success of Derryhouse (1902) and Trotter (1929) in elucidating ice flow trajectories based on erratics and till stratigraphy (see Hughes *et al.*, 1998 for a summary), geomorphological mapping

(e.g. Trotter *et al.*, 1929; King, 1976) and recognition of ice recessional features (e.g. Yorke *et al.*, 2007) has led to a partial understanding of the ice flow history in Northern England. However non-systematic, ‘patchy’ mapping of glacial landforms within the Tyne Gap (see Clark, *et al.*, 2004; Evans *et al.* 2005 for a summary) and poorly constrained chronological control has resulted in a ‘static’ model of ice flow behaviour throughout the Late Devensian.

The Tyne Gap’s location, directly east of major ice dispersal centres over the Solway Lowlands, Lake District and Scotland (Trotter, 1929), makes it an influential flow artery, capable of draining large volumes of ice from the western ice divide of the BIIS across to the east coast lowlands. This makes it a potential location for a major ice stream of the last BIIS (Beaumont, 1971; Bouledrou, *et al.*, 1988). The ability to elucidate whether or not the Tyne Gap hosted an ice stream has major implications for ice sheet configuration and stability not only throughout the Solway Lowlands and northern England but also down the English east coast (Raistrick, 1931; Catt, 1991).

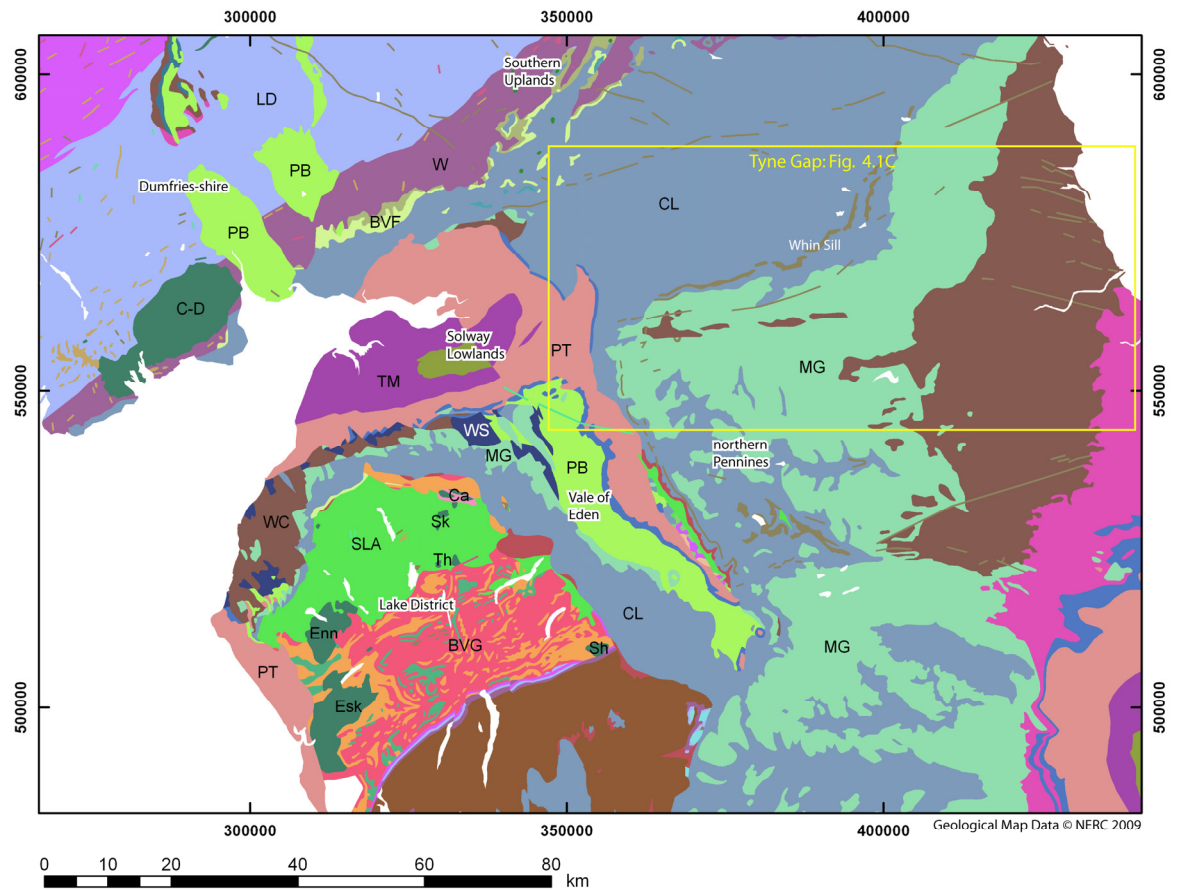
The aim of this paper is to present new glacial geomorphological and sedimentological data from the Tyne Gap in order to reconstruct the palaeo flow dynamics of this sector of the last BIIS and to assess what evidence there is for a palaeo-ice stream (e.g. Stokes and Clark, 1999, 2001). The paper focuses on reconstruction of former ice-flow trajectories and flow phasing through mapping of subglacial bedforms and the identification of their cross-cutting relationships. Quantification of elongation ratios allows relative flow velocities to be considered both spatially and temporally throughout the region, whilst deglacial features provide evidence of the nature of ice recession. Field-based assessments of stratigraphy, sedimentology and sediment provenance allow further evaluation of the ice flow dynamics in addition to providing information on former subglacial conditions at the ice-bed interface. In combination, these forms of data production allow a critical assessment of the Tyne Gap as a location for a palaeo-ice stream in the BIIS.

## 4.2 Study area

The Tyne Gap is a wide mountain pass located between the Scottish Southern Uplands to the north and the English Pennines to the south. It stretches 80 km from the Solway Lowlands in the west across to the mouth of the River Tyne in the east (Fig. 4.1). The Tyne Gap is underlain by Lower Carboniferous sedimentary rocks consisting of limestone, shale, sandstone and coal that dip south-southeastwards towards the edge of the northern Pennines (Fig. 4.2). These Carboniferous strata were deposited in a downwarping basin, the Northumberland Trough, bounded by the Southern Upland massif to the north and the north Pennines to the south. The northern margin of the north Pennines is delimited by the Stublick fault system which trends WSW-ENE. Resistant strata, including the Great Whin Sill, which trends SW-NE, result in strong west-east cuestas (Bouledrou *et al.* 1988). Strata of rhythmically bedded limestone, sandstone, shale and coal are exposed to the

## CHAPTER 4: A MAJOR ICE DRAINAGE PATHWAY - THE TYNE GAP

north and south of the Whin Sill, forming a series of scarps that are particularly prominent between Gilsland and the North Tyne Valley (Fig. 4.2). This strong geologic structure exhibited within the Tyne Gap has had a major influence on relief formation.



### Bedrock Geology:

Agglomerate in neck	Llanvirn and Arenig	Rhyolitic and trachytic lava and tuff
Andesitic and basaltic lavas and tuffs	Lower Lias	Rhyolitic lava
Andesitic lava and tuff	Lower Old Red Sandstone	Rhyolitic tuff, including ignimbrite
Ashgill	Lower Westphalian (coal measures)	Carboniferous Limestone Series
Ashgill and Caradoc	Ludlow	Triassic mudstones
Basal Conglomerate	Ludlow and Wenlock and Llandoverly	Tuff
Basalt and spilite	Magnesian Limestone	Tuff and agglomerate (mainly basaltic)
Basalt dolerite and camptonite	Mica-schists	Tuff (mainly andesitic)
Basalt, spilite and rhyolite	Middle Old Red Sandstone	Upper Greensand and Gault
Caradoc	MG Namurian (Millstone Grit Series)	Upper Lias
Diorite and allied intermediate types	PT Permian and Triassic sandstones	Upper Old Red Sandstone
Gabbro and allied types	PB Permian basal breccias	Wenlock
Granite, syenite and granophyre	Permian mudstones	Westphalian Coal Measures
Intermediate and basic rock	Porphyrite, lamprophyre	WS Westphalian and ?Stephanian
Limestone	Rhyolite, trachyte	
LD Llandoverly	Rhyolite, trachyte, felsite, elvans	

Figure 4.2: 1:625k bedrock geology map of northern England and the Scottish Borders. Granites: Sh = Shap, Esk = Eskdale, Enn = Ennerdale, Th = Threlkeld, Sk = Skiddaw, Ca = Carrock Fell, C-D = Criffel-Dalbeattie pluton. Extrusive igneous: BVG = Borrowdale Volcanic Group, BVF = Birrenwark Volcanic Formation. SLA = Skiddaw Slate Series. Note the Whin Sill orientated SW-NE in the Tyne Gap.

### 4.3 Previous research on the Late Devensian glaciation of the Tyne Gap

#### 4.3.1 *Subglacial bedform evidence for ice flow:*

Previous researchers have proposed that during the last glaciation ice flow from Scotland and the Lake District converged in the Tyne Gap and formed a major eastwards flowing routeway (Trotter, 1929; Johnson, 1952). The influx of ice from the Solway Lowlands, fed by Scottish and Lake District sources, provided the main supply of ice, although minor confluences have been identified at the junctions with the North Tyne, Allendales, Derwentdale, south Tynedale and Weardale (Clark, R. 1969). The north Pennines were characterised by local ice caps centred over Cross Fell, from which the Tees glacier emanated (Mitchell, 2007), and Cold Fell (Trotter, 1929; Vincent, 1969). Some evidence for palimpsest subglacial bedform signatures, such as the south-easterly orientated lineations in the North Tyne Valley, testify to cross-cutting, multi-phase ice flow (Clark, R. 2002; Livingstone *et al.*, 2008). The heavily lineated terrain, high relative elongation ratios (King, 1979) and convergent geometry, in the entrance to the Tyne Gap, have led some researchers to interpret the Tyne Gap as a zone of formerly streaming ice flow (Johnson, 1952; Bouledrou *et al.*, 1988; Everest *et al.*, 2005).

#### 4.3.2 *Erratics:*

Both Lake District (e.g. Borrowdale volcanic series, Carrock Fell gabbro, Permo-Triassic sandstone and Threlkeld grey quartz porphyry) and Southern Upland (e.g. Dalbeattie and Criffel granite, greywacke and Silurian grits) erratics are ubiquitous throughout the Tyne Gap. The erratic boundary associated with Lake District and Southern Upland provenances is indistinct, with a general southerly increase in Lake District erratics towards the north Pennines. This diffuse boundary is indicative of competing ice flows, with both Scottish and Lake District ice dispersal centres becoming dominant at different times (Lunn, 2004). From west to east the till in the Tyne gap changes from a red colour, associated with the Permo-Triassic sandstones of the Solway Lowlands and Edenside, to a grey colour, reflecting the Carboniferous rocks of the Tyne. This trend thereby records the eastwards decrease of Lake District erratics (Trotter, 1929) which in turn has been associated with a later phase of south-easterly, Scottish-sourced ice flow down the North Tyne Valley (Dwerryhouse, 1902).

#### 4.3.3 *Till stratigraphy in northern England:*

A tripartite succession of glacial deposits has been observed throughout large areas of the lowlands of northern England (Mackintosh, 1877; Trotter, 1929; Hollingworth, 1931; Smith and Francis,

1967; Hughes *et al.*, 1998; Huddart and Glasser, 2002). The succession has generally been attributed to Late Devensian glaciation, with glacial erosion being regarded as effective in removing evidence of older glaciations (see Chapter 2 for more details). The succession consists of a lower grey till and upper red till separated by stratified glaciofluvial or glaciolacustrine deposits. The top of the red till is often mottled, with Eyles and Sladen (1981) postulating that this was due to weathering. Till thickness throughout the Tyne Gap is variable, ranging from *ca.* 90 m in localised in-filled channels/valleys principally on the east coast (Hughes *et al.*, 1998) to extensive bare rock surfaces and veneers further inland. The tills within the tripartite sequence have been variably interpreted as lodgement tills with upper weathered horizons (Eyles and Sladen, 1981; Eyles *et al.* 1982), lodgement and melt out till continuums (Carruthers, 1953) and deformation tills (Hughes *et al.*, 1998). The stratified sediments separating the tills have almost universally been regarded as the product of subglacial meltwater deposition (Eyles *et al.* 1982; Clarke, B. G. *et al.*, 2008). Along the east coast of northern England two subglacial traction tills have been recognised, with the lower ‘Blackhall’ till characterised by a Southern Upland provenance and the upper ‘Horden’ till a Cheviot provenance (Davies *et al.*, 2009). Thus ice flow is initially thought to have flowed eastwards through the Tyne Gap (‘Blackhall’ till) before ice switched direction and instead became dominated by an ice lobe moving southwards down the eastern coast of Britain (‘Horden’ till) (Davies *et al.*, 2009).

#### 4.3.4 Deglaciation:

Two explanations have been proposed to explain deglaciation of the Tyne Gap. Firstly, several authors have suggested that deglaciation occurred by stagnation and *in situ* melting of the ice within the Tyne valley (Clark, 1969; Douglas, 1991; Mills and Holiday, 1998; Clark, R. 2002; Yorke *et al.*, 2007). Meltwater drainage deposited extensive sand and gravel deposits in the Tyne Valley (Yorke *et al.*, 2007), although Clark, R. (1969) has suggested that these deposits could have formed subglacially. Kamiform deposits are particularly extensive along the flanks of the present course of the Tyne (Yorke *et al.*, 2007) and there is evidence for a major lake (Glacial Lake Wear) which extended up the lower Tyne Valley (Teasdale and Hughes, 1999). Alternatively, Trotter (1929) proposed that deglaciation occurred by active frontal retreat with local glaciers of east Allendale, west Allendale and the south Tyne separating from westwards retreating ice in the Tyne Gap. As ice crossed the Tyne Gap, bifurcation of two ice lobes is surmised from two spreads of glaciofluvial deposits, one trending SW-NE with Lake District erratics and one trending NW-SE and containing Southern Upland erratics (Trotter, 1929).

## 4.4 Methods

#### 4.4.1 *Subglacial bedform mapping:*

Mapping of subglacial bedforms through the Tyne Gap involved the compilation of subglacial lineations, meltwater channels, eskers, ribbed moraine, hummocky terrain, glaciofluvial sediment accumulations and transverse ridges from NEXTMap 5 m resolution airborne Interferometric Synthetic Aperture Radar (IFSAR) imagery (cf. Livingstone *et al.*, 2008). The method follows the criteria of Clark (1997, 1999), Kleman *et al.* (2006) and Livingstone *et al.* (2008) in identifying discrete ‘flow sets’; defined as a collection of glacial features formed during the same flow phase and under the same conditions. Such flow sets are defined on the basis of conformity, length, parallelism and morphology (Clark, 1999). Cross-cutting relationships and superimposed bedforms have been used to construct a relative chronology, with flow sets assigned to distinct ice ‘flow phases’ (Boulton & Clark, 1990a, b; Clark, 1993, 1999; Kleman *et al.*, 2006). Six flow phases (with phase 6 being the oldest and phase 1 the youngest) were identified, with two sub-groups, which didn’t possess cross-cutting relationships (cf. Livingstone *et al.*, 2008). Several assumptions form the basis for much of the mapping technique (Kleman and Borgström, 1996): i) basal sliding requires a thawed bed; ii) lineations can only form if basal sliding occurs; iii) lineations are created in alignment to the local flow and perpendicular to the ice-surface contours at the time of creation; and iv) frozen bed conditions inhibit re-arrangement of the subglacial landscape.

Flow direction is identified by observation of the stoss and lee forms of drumlins and from provenance data, such as clast lithological analysis. Subglacial bedform length has been quantified by calculating elongation ratios (length/width) to assess variability in ice flow velocity both spatially and temporally (Stokes and Clark, 2001). Superficial (drift) (DiGMapGB-625 downloaded from the BGS website) and bedrock geology maps (BGS) have been overlaid onto the NEXTMap data in order to depict areas of bedrock moulded lineations and structural lithological influences. Higher resolution (1:50,000) superficial and geology maps have also been consulted. We also utilised the methodology of Kleman and Borgström (1996) in delineating a series of ‘fans’, which allowed the assessment of internal age chronologies, relative velocities, subglacial thermal regimes, stillstands, ice sheet configurations and potential ice streams. These were derived by combining ‘flowsets’ with other glacial landform assemblages, such as moraine, meltwater channels, glacial lakes and eskers (Kleman and Borgström, 1996).

#### 4.4.2 *Sedimentology and Stratigraphy:*

Borehole information and field sites at Hayden Bridge, Haltwhistle, Hexham and Willowford in the Tyne Gap (Fig. 4.1) are used to supplement the geomorphological mapping and to stratigraphically correlate ice flow events and identify reasons for the changing ice flow trajectories. Sediment stratigraphy and sedimentological analysis was undertaken at sections, allowing detailed analysis of

the depositional processes (Evans & Benn, 2004). Texture, sedimentary structure, colour, bed geometry, contacts and inclusions were all measured and logged, from which lithofacies were identified (Evans & Benn, 2004). Scaled section sketches were drawn of the larger exposures to assess sediment architecture and lateral lithofacies variability. Clast macrofabric analysis of the  $\alpha$ -axis orientation and dip (Benn, 2004), and investigations of clast lithology (Walden, 2004) supplement the geomorphological mapping of ice-flow pathways. Samples for thin section analysis were taken at Willowford to provide detailed micromorphological information on sediments at this site. Thin section analysis followed procedures outlined in van der Meer (1993), Menzies (2000), Carr (2004) and Hiemstra (2007). Borehole logs (at Haltwhistle and Hexham) provide a wider coverage for detailed stratigraphic correlation.

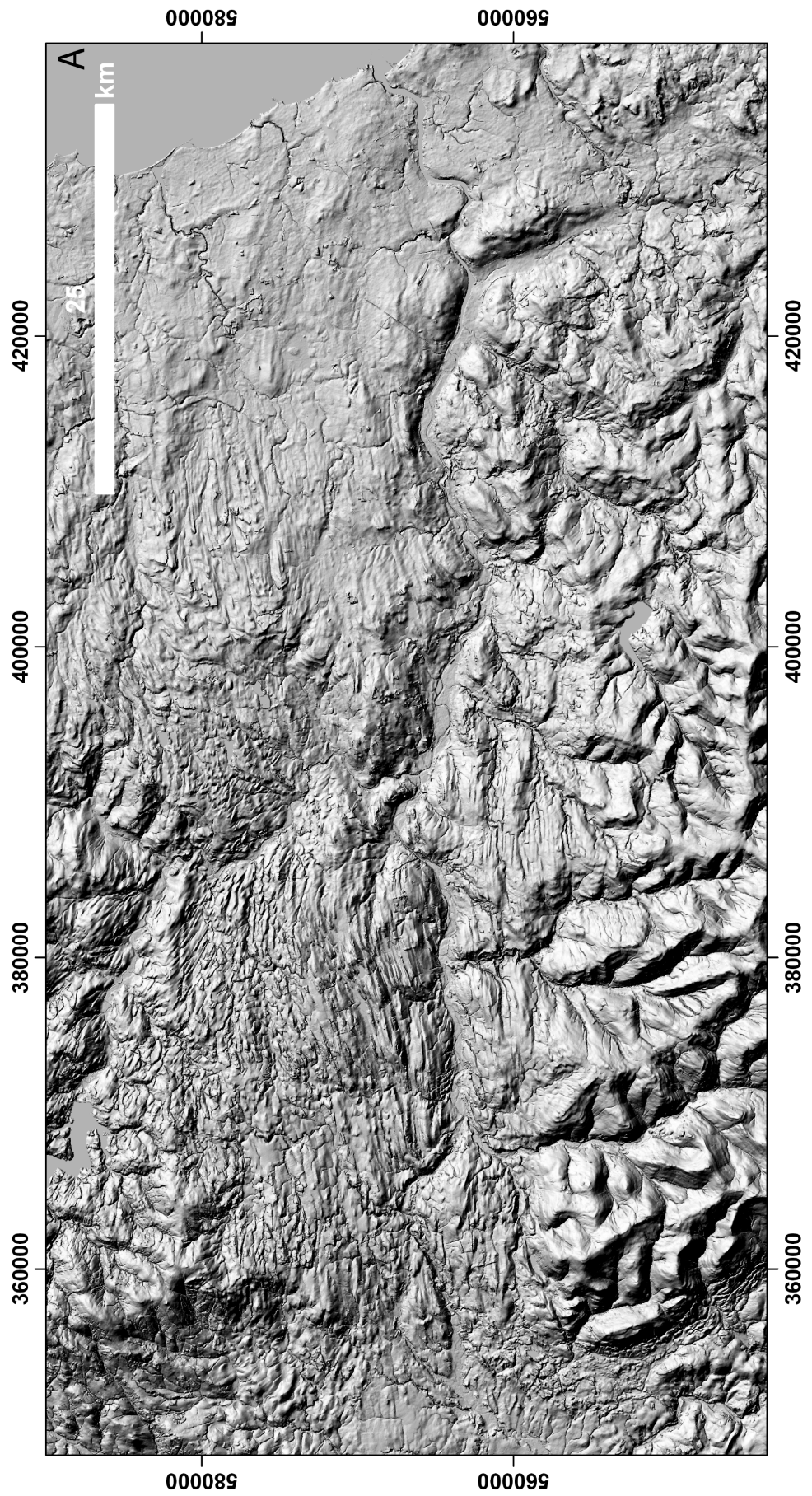
## 4.5 Results and Interpretation

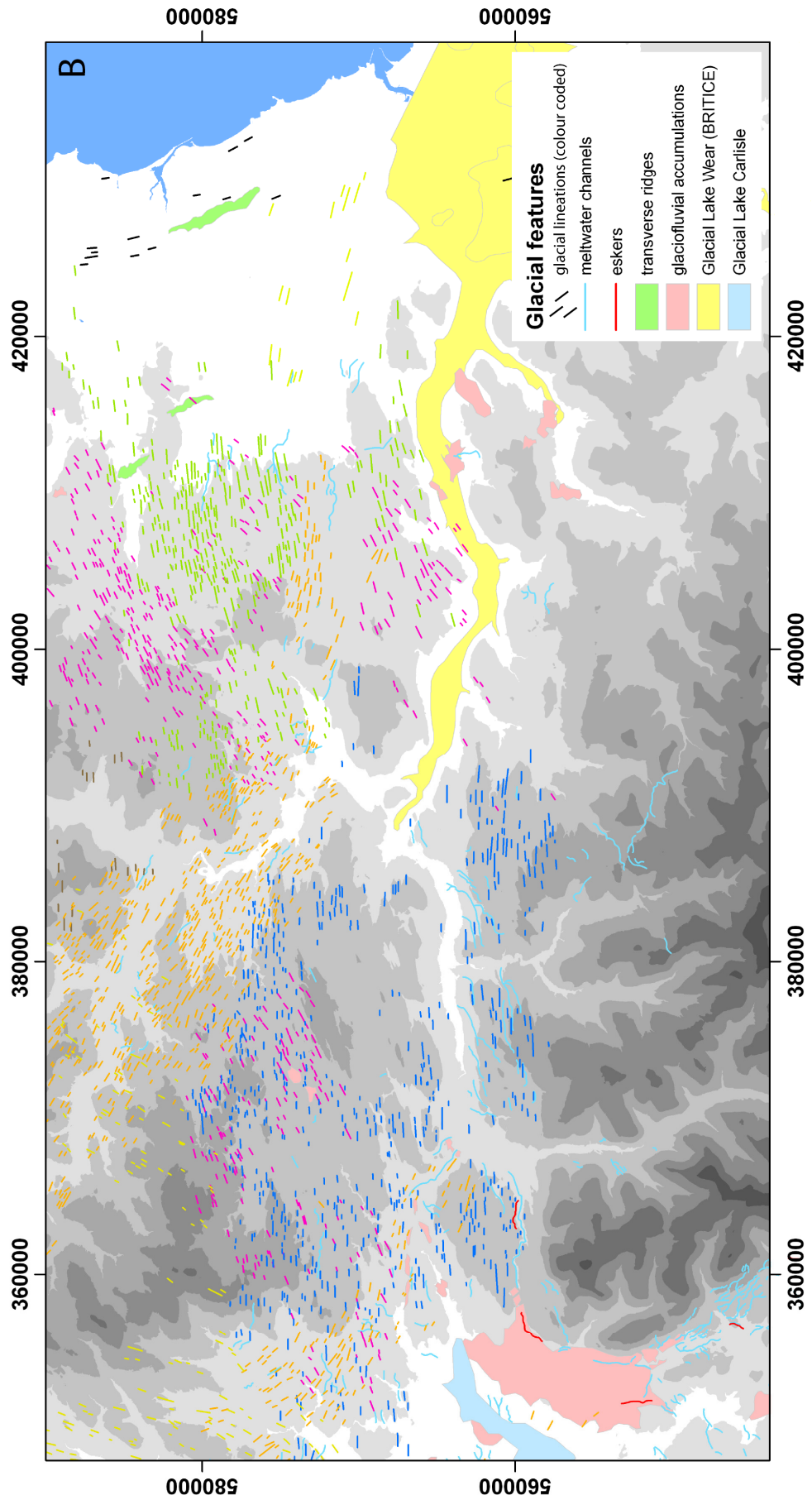
### 4.5.1 *Glacial Geomorphology:*

#### 4.5.1.1 *Streamlined subglacial bedforms*

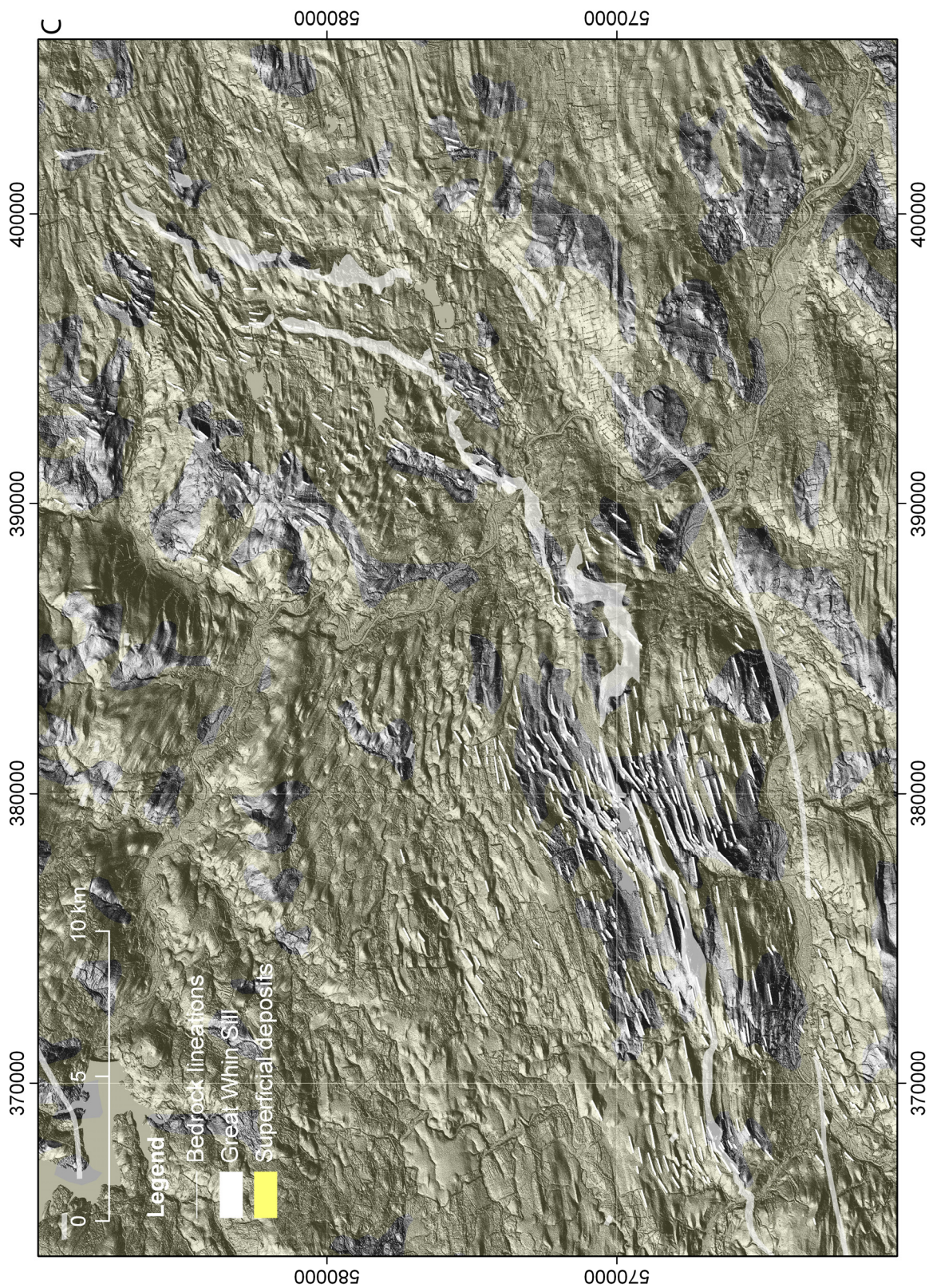
Lineations have been mapped throughout the Tyne Gap (Fig. 4.3a, b), with coverage ubiquitous throughout the lowlands (predominantly below the 400 m contour) (cf. Livingstone *et al.*, 2008). The subglacial lineations are up to 1500 m in length and comprise both bedrock and till forms. Results of the geomorphological mapping (Fig. 4.3) indicate that ice flow in the Tyne Gap during the last glaciation was predominantly towards the east. When split into flow sets, a series of ice flow shifts become apparent in the overprinting of subglacial bedforms (LT1-3: Chapter 2), suggesting a more dynamic pattern of ice flow (cf. Evans *et al.*, 2009). There are several features which typify the general ice-flow pattern throughout the Tyne Gap, including a central trunk zone, marked by a smooth corridor of streamlined terrain, and evidence of flow convergence from the Lake District and Southern Uplands (Fig. 4.3b). A major ‘set’ of lineations stretching south-east down the North Tyne valley and sourced from the central Southern Uplands is also apparent (LT4: Chapter 2). Towards the east coast the lineations become more subdued and eventually disappear with the exception of a small set of N-S orientated, isolated lineations (EC1: Chapter 2) (Fig. 4.3b). A similar scarcity of prominent lineations also exists in the Solway Lowlands, west of the Tyne Gap. The few bedforms present are subdued and hummocky and therefore more typical of Smith and Clark’s (2005) “ovoid forms”.



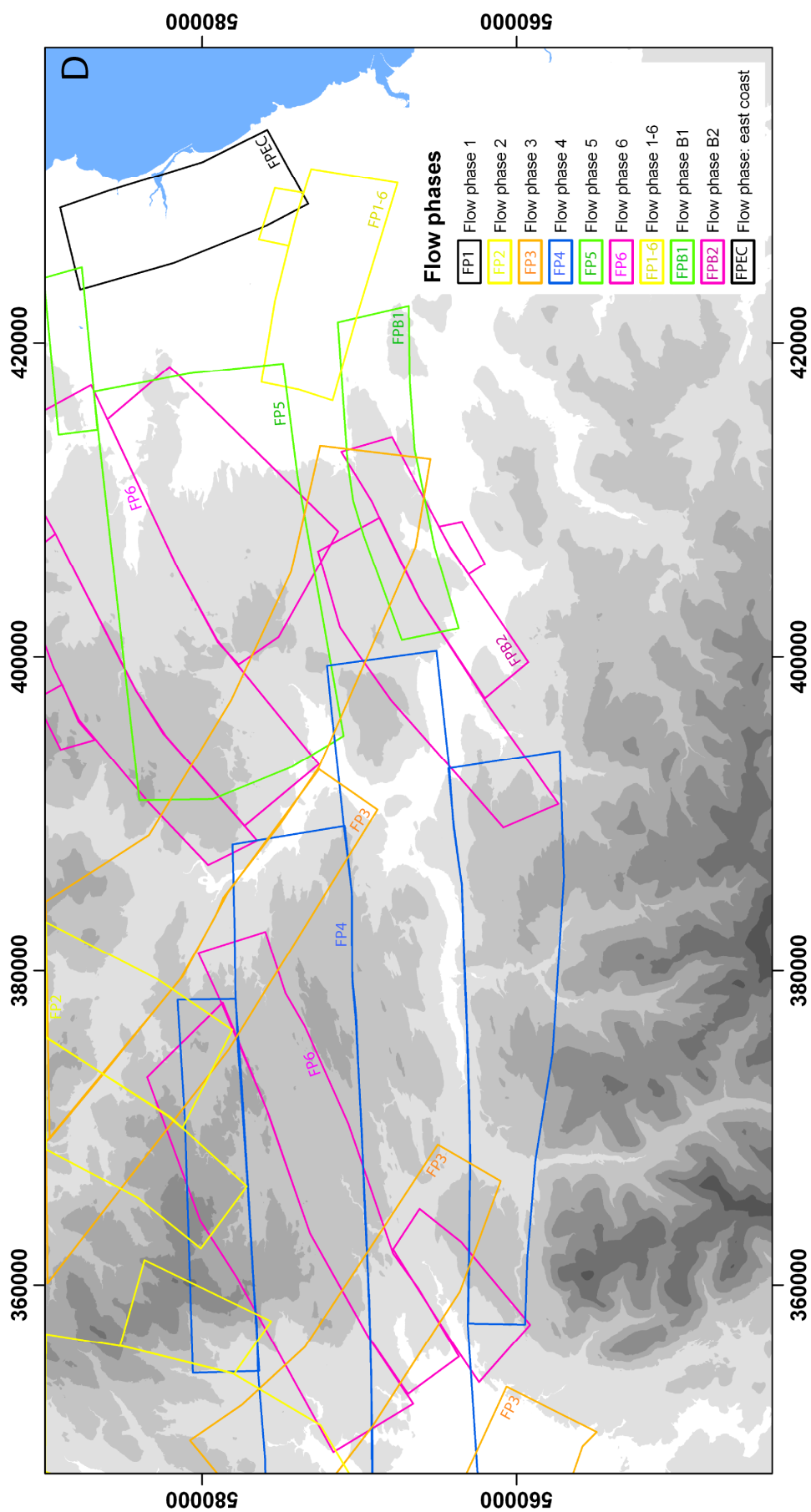












< Figure 4.3: (a) NEXTMap DEM of the heavily lineated Tyne Gap; (b) mapped subglacial bedforms (with lineations colour coded according to flow set); (c) bedrock structures and superficial deposits (DiGMapGB – 625 downloaded from the BGS); and (d) flow phases identified by cross-cutting relationships.

The subglacial bedforms mapped within the Tyne Gap are influenced by the arcuately exposed, Whin Sill dolerite, and south-easterly dipping, interbedded sandstone and limestone strata (Figs. 4.2 & 4.3c). This is characterised by visually striking streamlined terrain trending along the exposed bedrock geology (Fig. 4.3a). These have previously been mapped (Livingstone *et al.*, 2008) as subglacial bedforms pertaining to individual flow sets. However, preferential erosion along bedrock structures could theoretically produce two obliquely angled sets of structures produced during a single ice flow phase. 1:50,000 superficial deposits and bedrock geology maps (BGS) were used to assess which lineations were formed along exposed bedrock structures (Fig. 4.3c and also see section 3.4.1 for a thorough description of how bedrock and soft bedforms were differentiated). These lineations, although recognised as subglacial bedforms, are not included in the construction of flow sets here because of the ambiguity relating to which precise flow set they belong.

#### 4.5.1.2 *Relative chronology of flow sets:*

When organised into flow phases (Fig. 4.3d) based on cross-cutting relationships (cf. Evans *et al.*, 2009 for examples) initial ice flow was shown to originate from the south-west, i.e. the Lake District and associated northern Pennine tributaries (phase 6 (LT1: Chapter 2)). This was followed by a progressive shift in flow towards the east (phase 5 (LT2: Chapter 2) and 4 (LT3: Chapter 2), Fig. 4.3d) as the Scottish ice became increasingly influential. During this phase, valleys in the north Pennines continued to act as important secondary arteries (flow group B1 and B2, Fig. 4.3d). This period of ice movement is characterised by significant convergence of flow from the Lake District and Southern Uplands through the Solway Lowlands into the Tyne Gap. This increasing Scottish dominance resulted in the northern margin of the ice artery becoming breached by ice flowing down the North Tyne (phase 3 (LT4: Chapter 2), Fig. 4.3d) and is associated with Scottish ice flowing south-east into the Solway Lowlands. The final series of phases (phase 2 (LT5: Chapter 2) and 1 (LT7: Chapter 2), Fig. 4.3d) records a switch of flow towards a south-westerly direction, across the Tyne into the Solway Lowlands and out into the Irish Sea Basin. These latter phases of flow are topographically constrained (Fig. 4.3) suggesting that significant ice surface lowering and deglaciation occurred. The north-south orientated, east coast lineations (Fig. 4.3d) are related to a separate flow within the ice sheet. This, combined with their isolated position, makes it difficult to assign a flow phase to these features. However, as the Tyne Gap lineations of phases 6 and 5 wane towards the east coast (Fig. 4.3d) it can be inferred that these were removed by a subsequent north-south orientated flow along the east coast.

Elongation ratios of subglacial lineations are summarised in Table 4.1. They have been used to infer relative ice velocity throughout the Late Devensian (cf. Stokes and Clark, 1999). Phase 1 has not been included within the analysis, as the flow sets are not situated within the Tyne Gap.

	<b>Phase 2</b>	<b>Phase 3</b>	<b>Phase 4</b>	<b>Phase 5</b>	<b>Phase 6</b>
Count	267	745	859	324	421
Minimum	1.16	1.07	1.13	1	1
Maximum	4.11	4.94	8.85	9.91	9.6
Mean	2.19	2.15	3.26	2.72	2.23
SD	0.62	0.67	1.47	1.33	0.99

Table 4.1: Statistical summary of elongation ratios for ice flow phases 2-6 in the Tyne Gap (phase 1 not in Tyne Gap).

Table 4.1 shows that the general trend is of low average elongation ratios (between 2 and 4) throughout the region. The highest mean elongation ratio (3.26) is in flow phase 4 with the lowest (less than 2.2) in phases 2 and 3. All of the phases have very low elongation ratios. The maximum elongation ratio ranges between 4.11 for flow phase 2 up to 9.91 for flow phase 5. Throughout the flow series there is an initial lengthening of elongation ratios up to flow phase 4, followed by a decrease. This sudden decline to low mean elongations with little variation between lineations corresponds to a switch in flow, initially down the North Tyne valley and then SW into the Solway Lowlands. During flow phases 4 and 5 the elongation ratios are at their greatest mean, greatest maximum and greatest variance. There is also significant spatial variation in elongation ratios within individual flow sets, as seen by the range of elongation values (Table 4.1).

#### 4.5.1.3 *Meltwater channels*

Throughout the Tyne Gap there is a paucity of strongly aligned meltwater drainage channels (Fig. 4.3b). This is especially true for flow phases 6, 2 and 1. However, flow phases 3, 4 and 5 do show some evidence of associated meltwater drainage networks. Flow phase 3 has a NW-SE, flow trace-aligned series of channels. Flow phase 4 is associated with a series of parallel, ice marginal channels situated on the northern flanks of the Pennines and a smaller series of aligned channels within the Solway Lowlands, encroaching into the western end of the Tyne Gap (Fig. 4.3b). Flow phase 5 has a series of west-east aligned meltwater channels situated at the eastern edge of the flow group where the topography starts to become more subdued (Fig. 4.3b). There is a general absence of eskers throughout the Tyne Gap region.

The flow phases that are represented by abundant subglacial lineations but devoid of associated meltwater channels were likely formed as ‘synchronous’ fans (cf. Kleman and Borgström, 1996). This suggests that sheet flow within the interior of the ice sheet predominated at this time. The relationship of flow phases 3, 4 and 5 with meltwater channels suggests formation occurred primarily during ice recession (Kleman and Borgström, 1996). The absence of eskers during flow phase 3 (Fig. 4.5) suggests that subglacial drainage pathways did not develop, implying a ‘dry bed’

deglacial fan (Kleman and Borgström, 1996). The alignment of channels parallel to ice flow but in transverse swarms in the Solway Lowlands (Fig. 4.3b) during flow phase 4 infers a deglacial fan, thereby demonstrating a series of stillstand positions (Kleman and Borgström, 1996), with ice marginal/proglacial drainage occurring at the ice margin. The ice occupying the Solway Lowlands dammed drainage against the reverse slope of the Tyne Gap leading to breaching of the watershed, for example, at Gilsland (Trotter, 1929). Sequential deglaciation of the Tyne Gap occurred through successive ice surface lowering as evidenced by channels on the flanks of the Pennines. Ice-marginal drainage at the eastern edge of the Solway Lowlands corresponds to the final phase of west-east dominated flow through the Tyne Gap.

#### 4.5.1.4 *Moraines, glaciofluvial and glaciolacustrine deposits:*

Three NNW-SSE orientated ridges were identified within 18 km of the east coast (Fig. 4.3b): the 2.4 km long Wansbeck (NZ 113 850) drift barrier (Smythe, 1908, 1912); a 6 km long ridge at Cramlington (NZ 280 785); and a 2.8 km ridge east of Whalton (NZ 159 810). The ridges are transverse to the lineations associated with ice flow phase 5. The Cramlington ridge lies beyond the Tyne Gap flow sets, whilst the Wansbeck and Whalton ridges form a chain, indicating that they were formed concurrently. A major kame belt was identified in the Brampton region (Trotter, 1929; Huddart, 1970, 1981; Livingstone *et al.*, 2008), wrapped round the north-western edge of the Pennines and containing a series of ridges, flat topped hills and eskers of glaciofluvial sand and gravel (Fig. 4.3b). A superficial (drift) geology map (DiGMapGB-625 downloaded from the BGS website) has been overlaid onto the NEXTMap data in order to depict areas of glacial sand and gravel. A series of meltwater-dissected glaciofluvial sediments are identified along the Tyne Valley. Indeed, glaciofluvial deposits can be traced westwards to Gilsland, where they bifurcate, with one strand orientated NE-SW and another NW-SE (Fig. 4.3b). Two major glacial lakes were identified (Fig. 4.3b). Lake Wear (Smith, 1981; Teasdale and Hughes, 1999) was identified from the drift map and extended over large areas of County Durham and up the present day River Tyne. Lake Carlisle (Trotter, 1929; Huddart, 1970) was demarcated by a series of deltas positioned at different heights.

The geomorphic position of the transverse ridges, within, and at the end of, ice flow phase 5, in close association with a series of meltwater channels and kame deposits (Smythe, 1912), suggest that they are moraines. These moraines highlight the time-transgressive formation of ice flow phase 5 (Clark, 1999), marking a series of ice still-stands during recession. The absence of other defining features and the subdued topography associated with a faint N-S orientated flow set (east coast flow set, Fig. 4.3a) suggests that the ice flow phase was cross-cut by a southerly moving ice lobe impinging onto the east coast. This is supported by a two-tiered till stratigraphy; with the upper till consisting of a Cheviot provenance and the lower till a Tyne Valley provenance (Smythe, 1908; Davies *et al.*, 2009). Although the Cramlington ridge was formed within the N-S orientated east

coast flow set there is little evidence of overriding. However, lineations are very subdued in this region so it does not necessarily mean that the ridge postdates the southerly ice flow phase. The Brampton kame belt is a major ice-marginal depositional-centre (Trotter, 1929; Huddart, 1970, 1981) which started to form as ice receded through the Tyne Gap. The glaciofluvial deposits of the Tyne Valley also identified by Trotter (1929) and Yorke *et al.* (2007) record the former existence of a major drainage network that extended through the Tyne Gap.

#### 4.5.2 *Sedimentology and Stratigraphy:*

##### 4.5.2.1 *Willowford*

##### Description

The section at Willowford (NY 625 628) is located in a cutting of the River Irthing, just west of Gilsland (Fig. 4.1). The cutting is in the bottom 12.5 m of a 38 m high (160 m O. D.) lineation associated with flow phases 4 (SW-NE orientated) and 3 (NW-SE orientated); the surrounding area exhibits a series of cross-cutting landforms (flow phases 6, 4 and 3). There are 4 major sedimentary lithofacies associations in the section (Fig. 4.4).

Lithofacies association W1 (LFA W1) at the base of the section comprises up to 5.0 m of clast-supported pebble-gravel. The basal 1.8 m contains horizontally bedded, clast-supported rounded and sub-rounded gravel, which grades upwards into 1.3 m of massive, clast-supported boulder-gravel with occasional sand scours underneath individual boulders (Fig. 4.4). The boulders are up to 1 m in diameter and are angular/sub-rounded. The top 2 m of the sequence consists of horizontally stratified sand with a series of interbedded fills of massive pebble-gravel (Fig. 4.4). The channelled gravels have erosional contacts with the stratified sand, are rounded/sub-rounded and clast supported.

Lithofacies association W2 (LFA W2) comprises 0.9 m of laminated and slightly deformed sand and silt overlying LFA W1 (Fig. 4.5a, b). The basal 0.3 m consists of upwards fining coarse sand and laminated silt and clay rhythmites (Fig. 4.7b) with occasional dropstones up to 5 cm in diameter. Deformation is manifest as centimetre-scale contortions and wavy laminae (Fig. 4.5a, b) which show no sign of internal, penecontemporaneous soft-sediment deformation. The overlying 0.6 m of LFW2 are silt laminations dipping towards the SW at an angle of 18° (Fig. 4.4). These grade up into similarly orientated medium sand laminations, which contain attenuated clay pods forming flame structures, dipping, wavy laminae and contortions typical of soft sediment deformation (Fig. 4.5b).

Lithofacies association W3 (LFA W3) is characterised by dipping interbeds, <20 cm thick (Fig. 4.7c, d), of diamicton and sand, separated from the underlying LFA W2 by an erosional, undulatory



boundary (Fig. 4.4). The diamictos are massive, reddish (5 YR 3/3) and brownish-black (10 YR 3/1) coloured and matrix supported, with variable amounts of sub-rounded clasts (Fig. 4.5c,d). Loaded contacts and convolute bedding (Fig. 4.5c) suggest that soft sediment deformation persisted during deposition. Sand from LFA W2 has been sheared up and incorporated within the sequence (Fig. 4.5c,d).

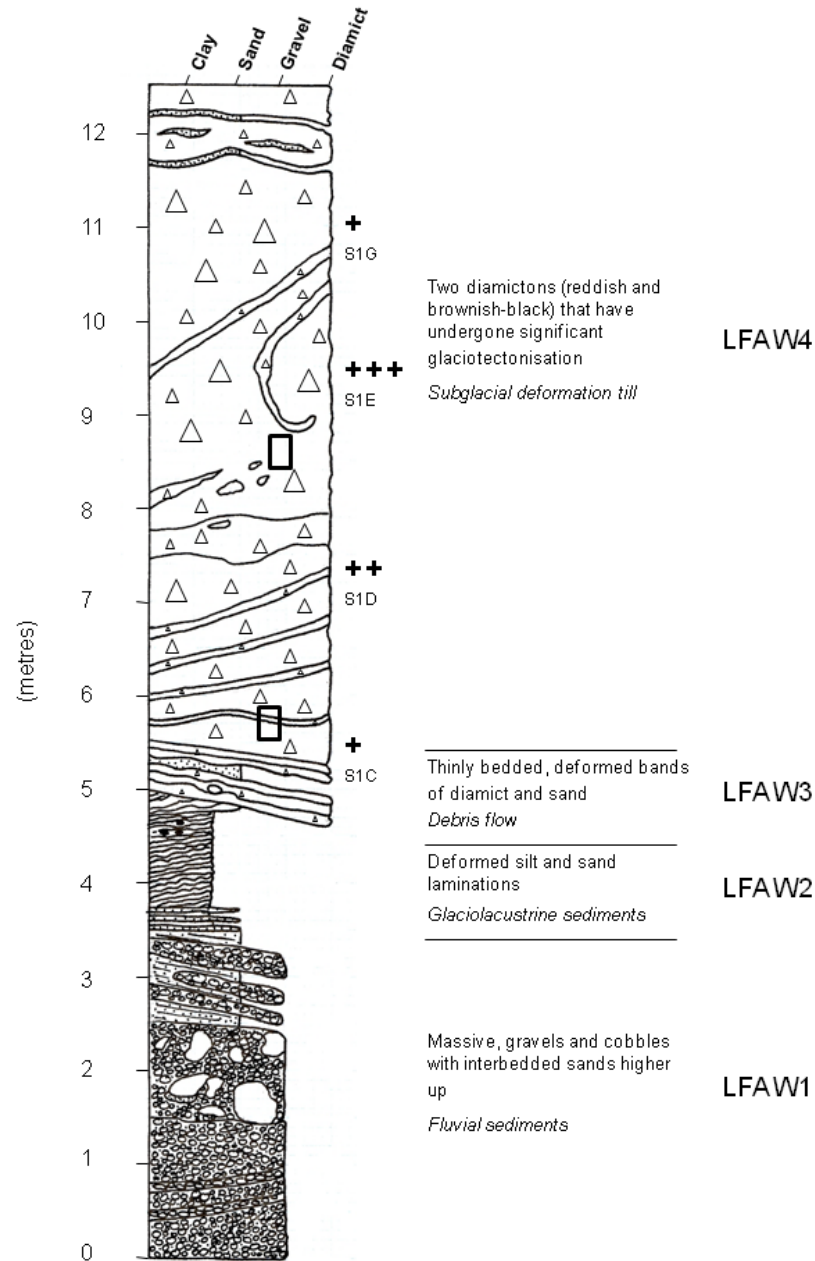
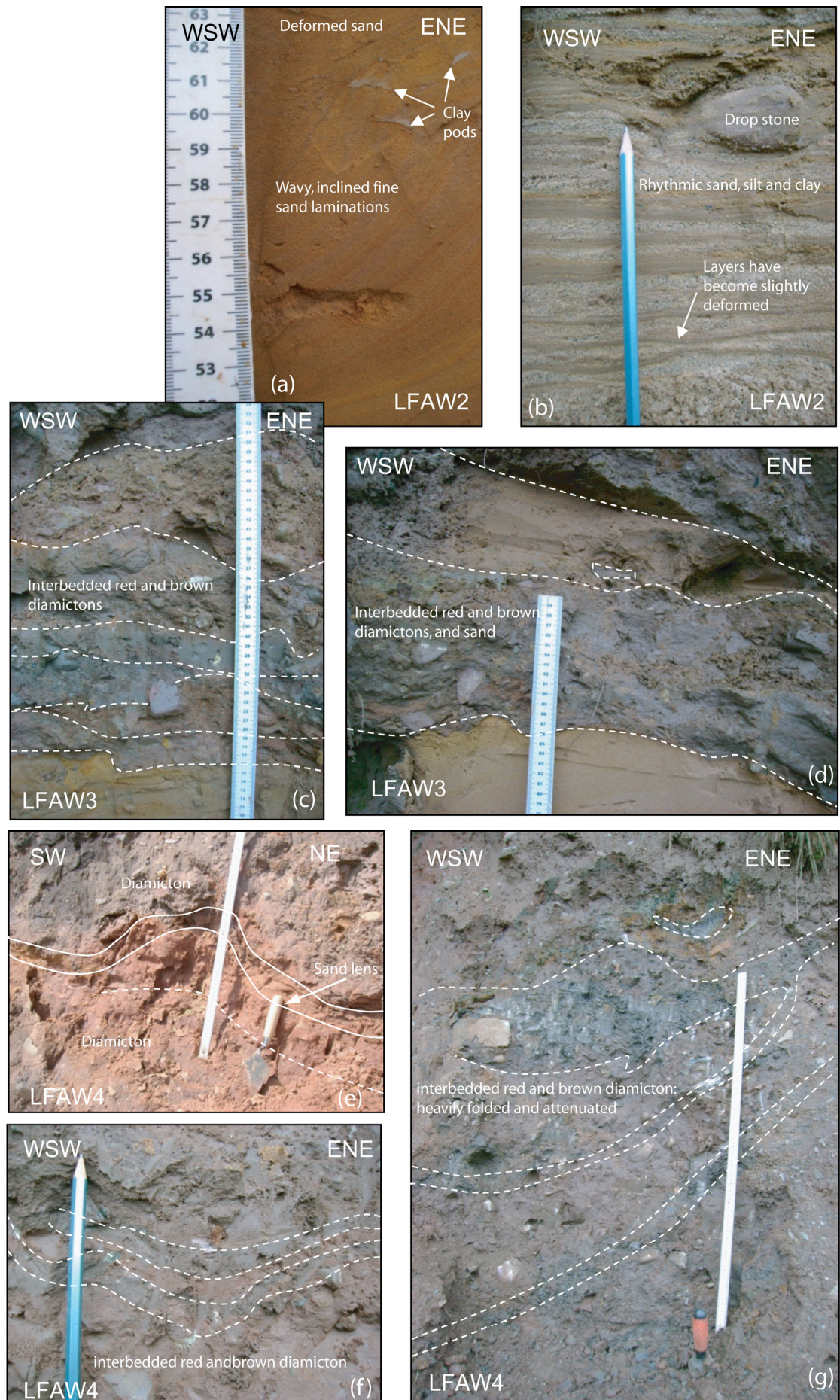


Figure 4.4: Composite stratigraphic log from Willowford field site (lithofacies codes from Benn and Evans, 1998). Boxes indicate thin sections and crosses refer to clast fabrics taken from the sections. Multiple fabrics have been taken laterally across some field sections.





< Figure 4.5: (a) and (b): LFA W2: examples of slight deformation structures within sands and silts interpreted as glaciolacustrine deposits. (c) and (d): LFA W3: examples of thinly bedded, deformed diamictons and sands interpreted as debris flow tills. (e) – (g): LFA W4: Glaciotectonised diamictons (reddish and brownish-black), with interbedded sand layers near the top of the sequence interpreted as a subglacial deformation till.

Lithofacies association W4 (LFA W4) comprises 6.8 m of inter-bedded reddish and brownish-black diamictons (Fig. 4.5e-g) which have been heavily deformed, as evident by a series of folds and attenuations (Fig. 4.4). The reddish (5 YR 3/3) diamicton is massive, matrix-supported, has a silty-sandy texture which becomes increasingly sandy up-sequence, and contains abundant sub-rounded to sub-angular clasts (Fig. 4.5g). In the upper 3 m of LFA W4 the reddish colouration locally turns dull yellowish-brown (10 YR 4/3). The brownish-black (10 YR 3/1) diamicton is predominantly massive, and contains some sand lenses higher up in the sequence (Fig. 4.5e). It is matrix-supported, friable, with a clayey-silty texture and has a moderate density of sub-rounded to sub-angular clasts. In places the brownish-black diamicton contains pockets of angular clasts. Both diamictons have a high frequency of striae, with the angle of orientation parallel to the a-axis of the clasts. The reddish diamicton dominates throughout LFA W4, with the brownish-black diamicton interfingering as centimetre scale beds (Fig. 4.5f, g). Centimetre thick, wavy, roughly horizontal, sand laminations occur between and within the diamicton beds in the top 3 m and bottom 1 m of LFA W4 (Fig. 4.5e and see micromorphology below). LFA W4 lacks major shear planes, but does contain folds and attenuated beds dipping west-southwest (Fig. 4.5g). North-east of the diamictons a sand unit laterally contiguous with LFA W4 was observed. This could be a sand fold or raft. Two thin sections collected from LFA W4 (see Fig. 4.6) display a number of rotational structures often associated with a series of lineations and grain-stacking arrangements. Two diamicton domains cutting obliquely down towards the ENE with folded, convoluted and diffuse boundaries have also been noted (Fig. 4.8b), whilst one of the thin sections contains a horizontally aligned and laminated soft-sediment stringer which had been subjected to extensive deformation (Fig. 4.6a) (series of faults, folds and marble textured fine clay and silt).

Clast macrofabrics from both the reddish and brownish-black diamictons of LFA W4 are shown in Fig. 4.7 (with locations shown on Fig. 4.4).  $S_1$  eigenvalues range between 0.44 and 0.74, with fabrics mirroring the ice flow direction (SW-NE) in the upper half of the diamicton and orientated at right-angles in the lower half of the diamicton. Clast lithological analysis ( $n = 300$ ) was carried out on the gravels in LFA W1 and both the reddish and brownish-black diamictons in LFA W3 and 4 (Fig. 4.8). The gravels are dominated by Carboniferous limestone (the local bedrock), quartzitic sandstone, greywacke and coal. The red and grey diamictons contain similar lithologies (Fig. 4.8). Both have mixed clast lithologies including andesites, rhyolites and slate from the Lake District, local Carboniferous limestone, Scottish greywacke, Criffel and Dalbeattie granite and metamorphics, and Solway Lowland (Permo-Triassic sandstone) rocks (Fig. 4.8).

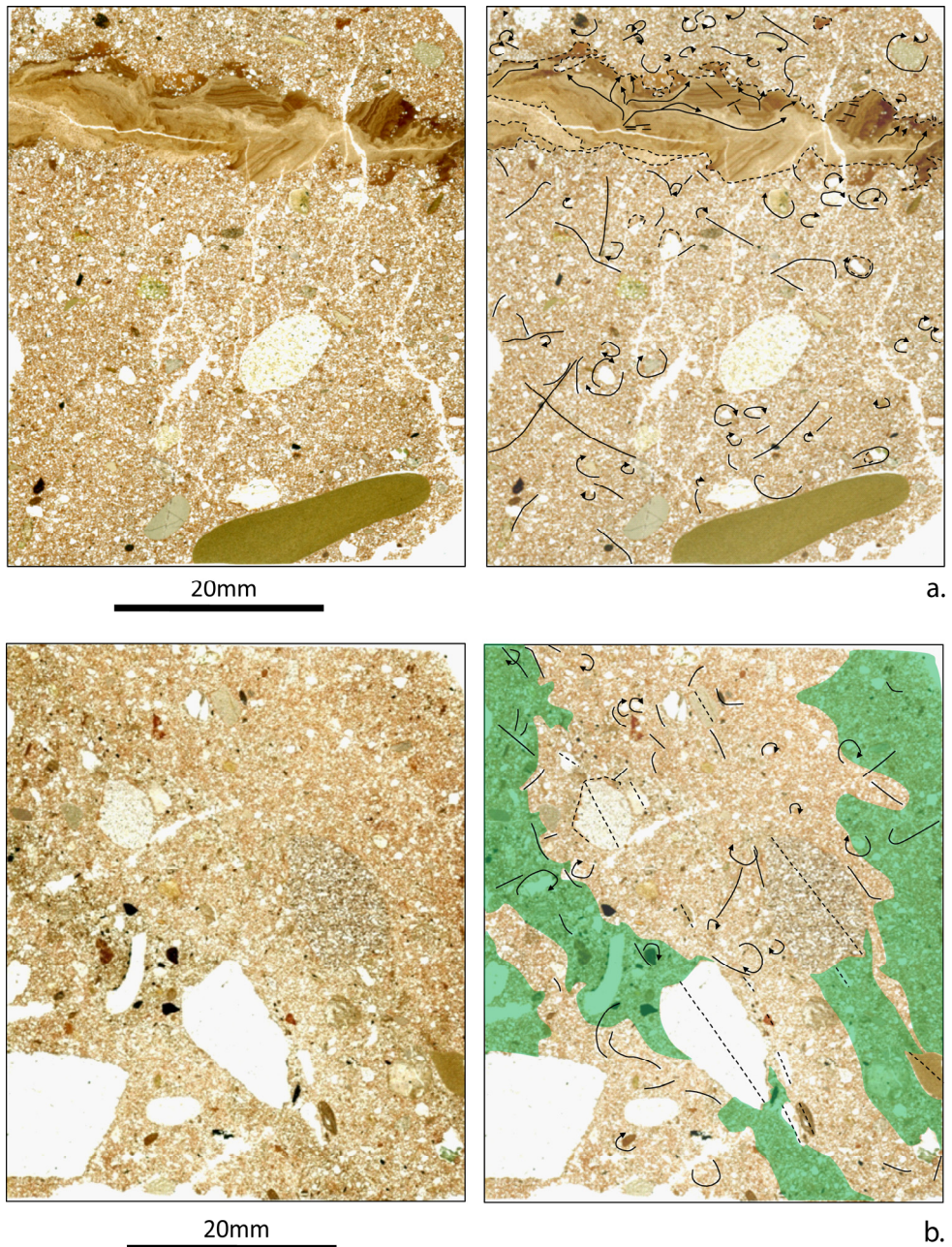


Figure 4.6: Thin sections from Willowford (see Fig. 4.4 for location: 4.6a is the lower thin section and 4.6b is the upper thin section). All paired-images are scans of the original thin sections, while the scale bar is located at the bottom of each original image. In all images arrows indicate turbate structures, dotted lines around grains delineate sediment caps and lines demarcate lineations/grain stacking. In 6a the fine-grained laminae is marked by dotted lines, while within this domain arrows indicate water escape structures and lines faults. In 4.6b the green blocks delineate the two diamicton domains while dotted lines superimposed on grains illustrate their preferred orientation. The thin sections are orientated (from left to right) WSW to ENE, with the main regional ice flow orientated SW-NE. Image (a) is characterised by large numbers of turbate and lineation structures within the diamicton and a horizontally orientated soft sediment laminae. This laminae is heavily deformed with a sheared upper boundary, and evidence for extensive fluidisation. Image (b) is characterised by two distinct diamicton domains, orientated grains and evidence of both turbate and lineation structures.

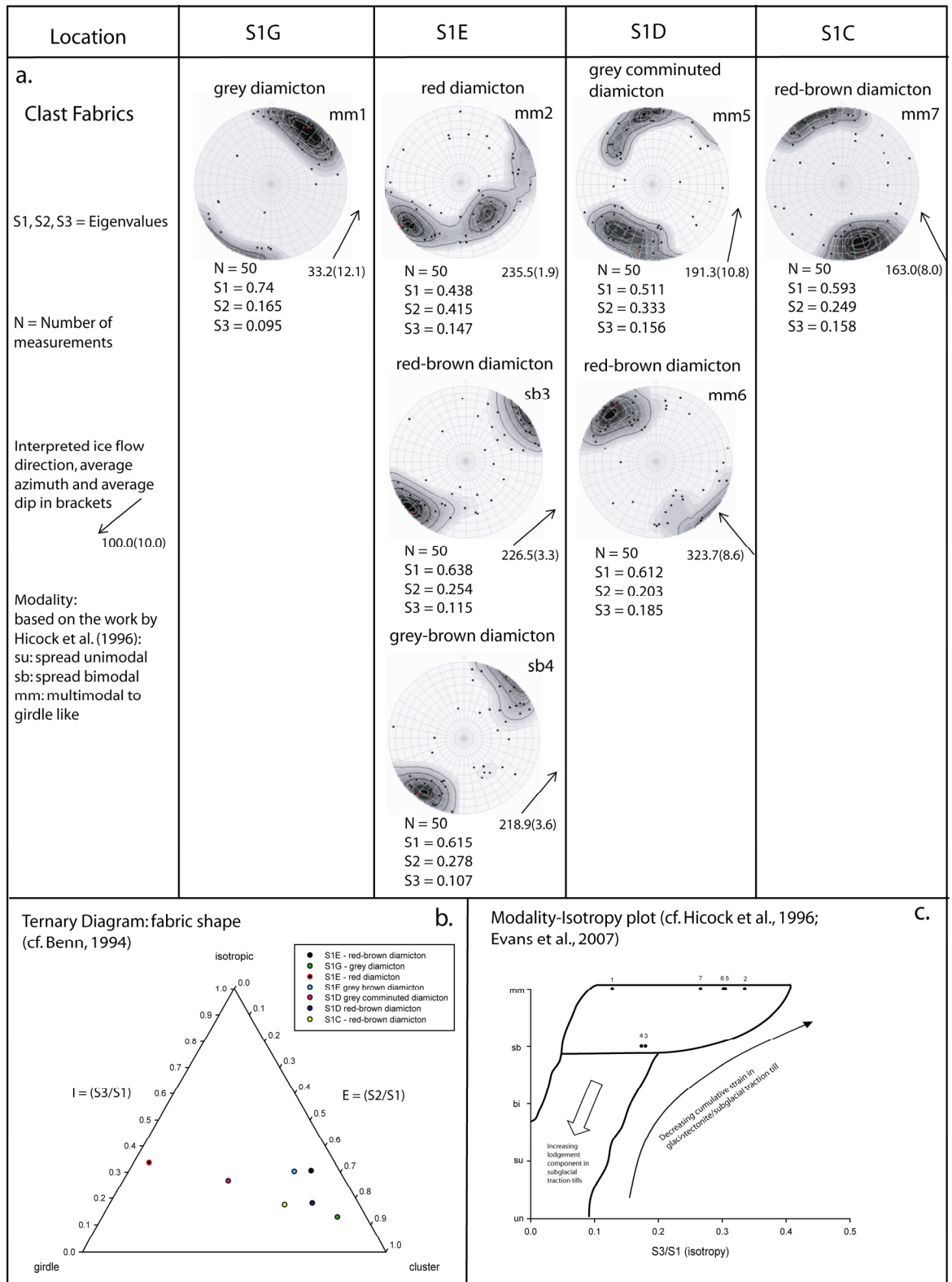


Figure 4.7: (a) Clast macrofabric data for the two tills within LFA W4, at Willowford (samples taken from 50 clasts from locations shown on Figure 4.7). (b) Ternary Diagram showing clast fabric shape (cf. Benn, 1994). (c) Modality-Isotropy plot (cf. Hicock et al, 1996).

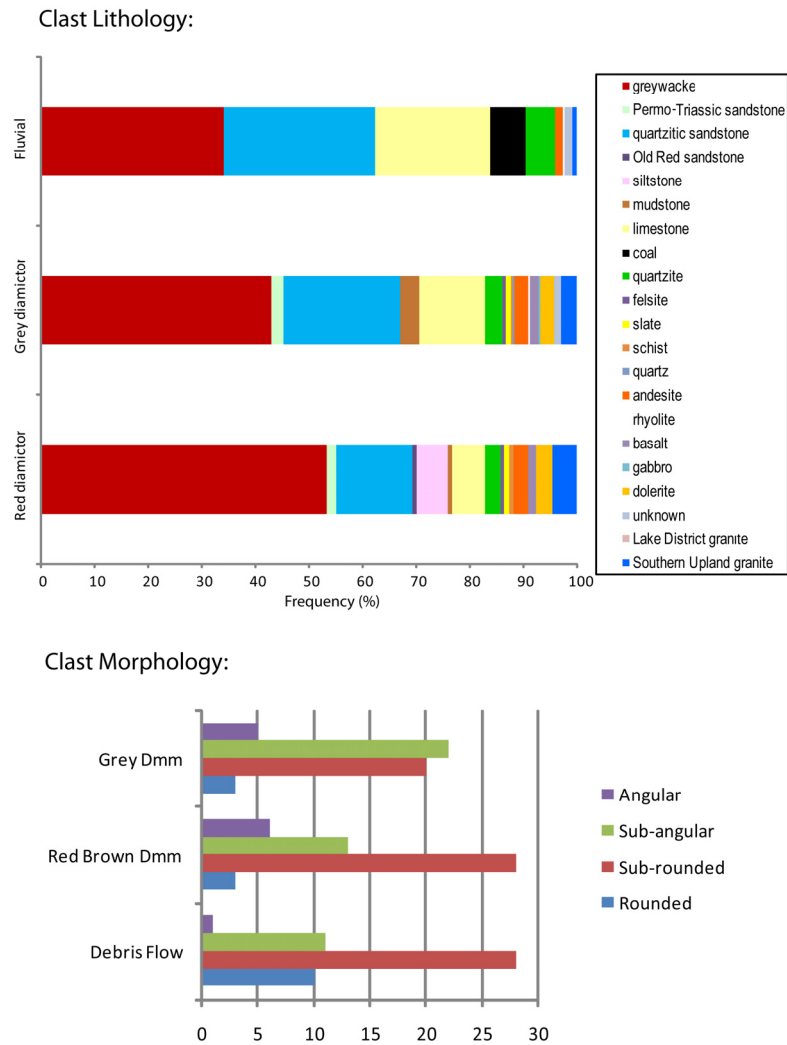


Figure 4.8: Provenance data at Willowford: clast lithologies (n = 300) collected from the red and grey diamicton and fluvial gravels; and clast morphologies (samples taken from 50 clasts).

#### Interpretation:

The massive cobble-gravel with interbedded sand towards the top of LFA W1 is indicative of a fluctuating glaciofluvial depositional regime. These sediments were deposited in an increasingly higher energy system which was capable of transporting clasts up to 1 m in diameter and characterised by a series of gravel bars typical of a proglacial braided river network (Miall, 1977). Given the absence of any diamicton this indicates that ice had not yet over-ridden the Tyne Gap area. The overlying laminated silts and sands with dropstones (LFA W2) are interpreted as glaciolacustrine deposits. A slight coarsening of the silts and sands up-sequence may reflect increasingly ice proximal conditions or just increased discharge. The small-scale deformation structures were probably imparted by a later stage of glacial over-riding. The overall fining-upward sequence represented by LFA W1 and W2 is a depositional record of fluvial drainage dammed by encroaching glacier ice. The arrival of the ice in the drainage basin is recorded directly by the



thinly bedded and deformed sands and diamicton with loaded boundaries of LFA W3. LFA W3 was deposited by intermittent debris flows as suggested by the thin diamicton beds (Lawson, 1979) and sheared sands from LFA W2 (Lawson, 1981). Load structures and deformation in LFA W3 suggest high water contents and concomitant soft sediment deformation (Lawson, 1981). The presence of far-travelled erratics indicates that these sediments were related to glacial advance.

The reddish and brownish-black diamictons of LFA W4 are interpreted as subglacial traction tills (*sensu* Evans *et al.*, 2006). Evidence for this includes: (a) stones that are commonly striated and have a far-travelled provenance; (b) Intercalated diamictons exhibiting a series of dipping fold structures, attenuations and boudins typical of deformation tills; (c) very compact, monolithological, clast dominated, pockets of brownish-black diamicton interpreted as rafts of bedrock that have been crushed and rafted up into the diamicton (e.g. Hiemstra *et al.*, 2007); (d) microstructures of both of both ductile (skelsepic fabric, turbates) and brittle (lineations, grain stacking, fissility) deformation typical of a high stress polyphase till (van der Meer, 1993; Menzies, 2000; Menzies *et al.*, 2006); and (e) macro-fabric orientations and  $S_1$  eigenvalues in the upper sections which are consistent with the inferred direction of ice flow (Benn, 1995; Evans, 2000).

The range of  $S_1$  eigenvalues in LFA W4 (Fig. 4.7) indicates that the mode of formation was polygenetic, thus suggesting that the tills are hybrids (Hicock and Fuller, 1995; Nelson *et al.*, 2005). The horizontal sand stringers seen both micro- and macroscopically are interpreted to have formed in a low energy subglacial environment, possibly during decoupling of the basal ice and the development of a thin water film (cf. Piotrowski and Tulaczyk, 1999; Piotrowski *et al.*, 2001, 2002, 2004, 2006). The fact that the stringer survives suggests that deformation was not pervasive throughout the unit (Piotrowski, *et al.*, 2001). Indeed, incomplete homogenisation of the two tills indicates low strains (Khatwa and Tulaczyk, 2001), thus suggesting that the till is immature. The mixed provenances displayed by both the reddish and brownish-black diamicton could have resulted from subglacial cannibalisation of pre-existing sediment, possibly deposited from previous glacial episodes or by way of the shifting dominance of ice dispersal centres in the Lake District and Scotland (Lunn, 2004). However, the brownish-black diamicton does contain a higher percentage of local Carboniferous lithologies which when interpreted in conjunction with the two distinctive colours suggests different flow paths or timings of flow.

Throughout Cumbria and Northumberland the general glacial stratigraphy has been observed to be a tripartite sequence consisting of a lower grey till, middle sands and upper red till (Huddart and Glasser, 2002). This would suggest that the brownish-black diamicton at Willowford was deposited first, followed by glaciotectionisation and deformation during a second phase of deposition (reddish till). However, the underlying debris flow unit (LFA W3) consists of both reddish and brownish-black diamicton and, coupled with the absence of a distinctive lower brownish-black till, precludes

an interpretation of multiple phases of glacial deposition at Willowford (although multiple phases of ice flow did occur).

#### 4.5.2.2 Hayden Bridge

##### Description:

Hayden Bridge (NY 844 638) is located on the southern flank of the South Tyne Valley (Fig. 4.1). The site is located at an elevation of 160 m O. D., 95 m above the South Tyne River (65 m O. D.), and is in a region containing flow phase 4 (west-east orientated) lineations. There are three major sedimentary associations identified from 2 sites and borehole data (LFA HB1-3; Fig. 4.9 and 4.10a-d).

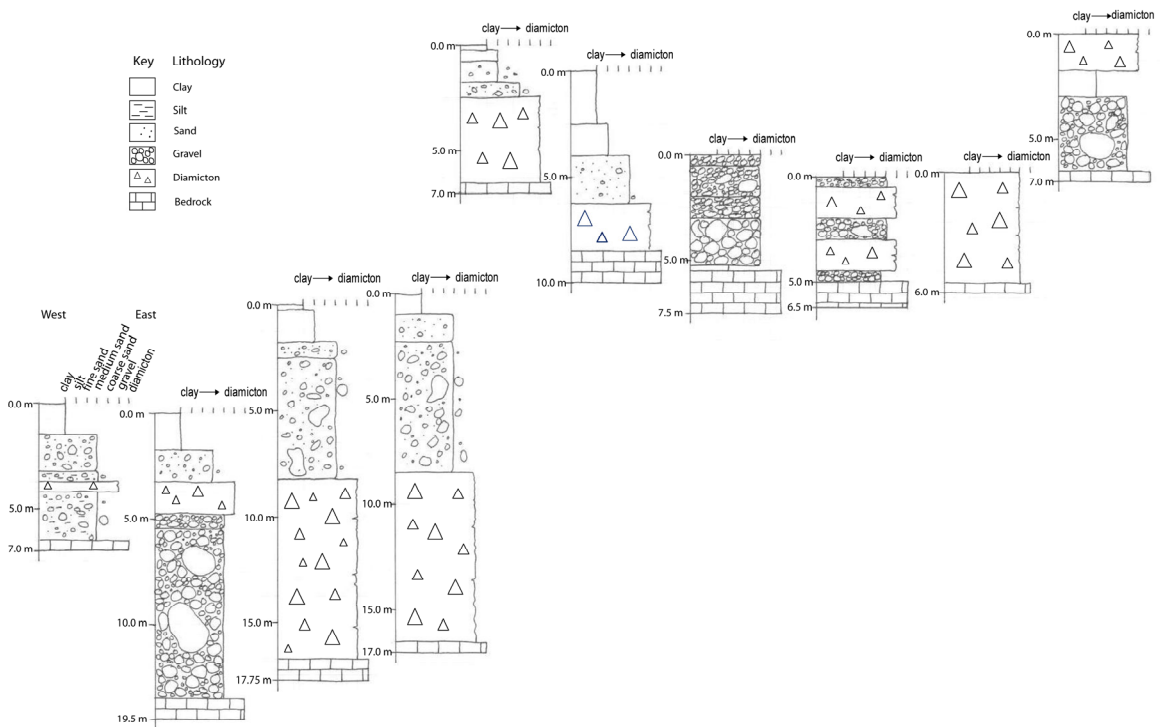


Figure 4.9: Borehole logs (BGS) showing stratigraphic sections and borehole locations at Hayden Bridge field site.

Lithofacies association HB1 (LFA HB1) comprises >6 m of matrix-supported, brownish-black (7.5 YR 3/1) diamicton (Fig. 4.10b) containing sand lenses and rounded/sub-rounded clasts of predominantly local lithologies (sandstone, limestone and dolerite), although there are some far-travelled erratics such as St Bees sandstone, greywacke and green andesite. Rare striations were observed on the clasts. The sand lenses predominate in the upper 3 m and the upper boundary with LFA HB2 is erosional. This diamicton is situated either directly on top of the bedrock or on an intervening unit of sand and gravel.



Lithofacies association HB2 (LFA HB2) is a discontinuous series of boulders, gravel and sand (Fig. 4.10a). At site 1 this is *ca.* 2 m thick near the base of the section and is composed of cobbles and boulders (up to 0.5 m in diameter), overlain by stratified and massive gravels and sands. This implies a general sequence of fining that grades up into LFA HB3. At site 2 the gravel and sand of LFA HB2 is massive throughout.

Lithofacies association HB3 (LFA HB3) consists of *ca.* 3 m of laterally discontinuous, stratified silts and sands fining upwards into laminated clay (Fig. 4.10c, d). The laminated clay sequence contains a series of centimetre-scale sand lenses. The very top of the sequence coarsens into sandier beds which grade into Lithofacies HB 4, comprising up to 5 m of massive, matrix supported sand and coarse gravel (Fig. 4.10c).



Figure 4.10: Photographs of lithofacies exposed at Hayden Bridge: (a) gravels and sands of unit 2; (b) diamicton of unit 1; (c) silts and sands which fine up into clay laminations. These are overlaid by sand and gravel deposits; (d) silts and sands which fine up into clay laminations.

Interpretation:

LFA HB1 is interpreted as a subglacial till based on its massive appearance, the occurrence of striations and far-travelled erratic content. The overlying matrix-supported boulder, gravel and sand deposits (LFA HB2) are interpreted as glaciofluvial in origin due to the abrupt lateral and vertical changes in particle size and geometry, which are characteristic of the fluctuating discharges of proglacial streams (e.g. Miall 1977; Marren, 2001). The discontinuous diamicton unit situated between sand and gravel lithofacies could relate to either debris flow deposits or a subglacial till associated with an oscillating ice front. LFA HB 2 therefore represents a proglacial, braided drainage network that drained eastwards along the present course of the South Tyne River. The general fining upwards trend within the deposits suggests an increasingly ice distal environment (e.g. Marren, 2001) associated with ice recession westwards over the Tyne Gap watershed. As ice retreated westwards a series of meltwater channels drained into the main Tyne Valley (e.g. the Gilsland meltwater channel). To the west of Gilsland the deposits are split into a SW-NE ‘train’ dominated by Lake District erratics and a NW-SE ‘train’ by a Scottish provenance (Trotter, 1929), thus indicating a decoupling of two previously convergent ice flows.

The stratified silts and sands grading upwards into laminated clays (LFA HB3) are interpreted as glaciolacustrine sediments deposited in a lake formed in the Tyne Valley.

4.5.2.3 *Borehole logs*

Haltwhistle Bypass:

Logs of site investigation boreholes have been drilled along the Haltwhistle Bypass (NY 705 632) at *ca.* 130 m O. D. on the flank of the South Tyne River just to the east of Gilsland meltwater channel (Fig. 4.1). Haltwhistle Bypass is in a region of flow phase 6 and 4 lineations, with a series of meltwater channels, orientated towards the Tyne Valley, dissecting the landscape to the west and south.

There is a four-fold division of sediments within the borehole logs (Fig. 4.11) drilled higher up on the flanks of the South Tyne Valley. The lowest lithofacies comprises a basal (*ca.* 10 m thick) diamicton which changes upwards in colour from dark brown to red-brown, the boundary being marked in places by thin clay/sand intercalations. This lithofacies is followed by boulders, gravel and sand which vary in their lateral extent and have a range of sedimentological characteristics, ranging from dense grey sand with many cobbles and boulders to brown silty-sand. The upper diamicton is less than 2 m thick and has a mottled appearance. The upper sands and gravels, like the lower lithofacies, also experience progressive upwards fining. This lithofacies is absent within many of the boreholes. The borehole logs drilled at lower elevations Tyne (Fig. 4.11) are composed

almost entirely of sand and gravel deposits resting directly on the bedrock, with the exception of borehole 238, which contains a thin orange-brown, mottled diamicton.

The sandwiching of sands and gravels between diamictons, as displayed at Haltwhistle Bypass, is typical of many sites in northern England (cf. Huddart and Glasser, 2002). The generally accepted theory on the mode of formation developed from the ideas of Boulton (1977) in Wales and entails advance and decay relating to a single glaciation (Goodchild, 1875; Eyles and Sladen, 1981; Huddart and Glasser, 2002). However, information gathered from borehole logs is not enough to make a firm classification of the diamicton other than its glacial origin. The upper diamicton is mottled, which could indicate that it has been weathered, as is typical of diamictons along the east coast (Eyles and Sladen, 1981). The sands and gravels which stratigraphically overlie diamictons and have variable textures are interpreted as glaciofluvial outwash. The upwards fining within both the sand and gravel lithofacies implies an increasingly ice-distal environment, while absence of diamicton balls within the sequence entails proglacial drainage, with the boulder and cobbles indicative of a high energy glaciofluvial system (Miall, 1977).

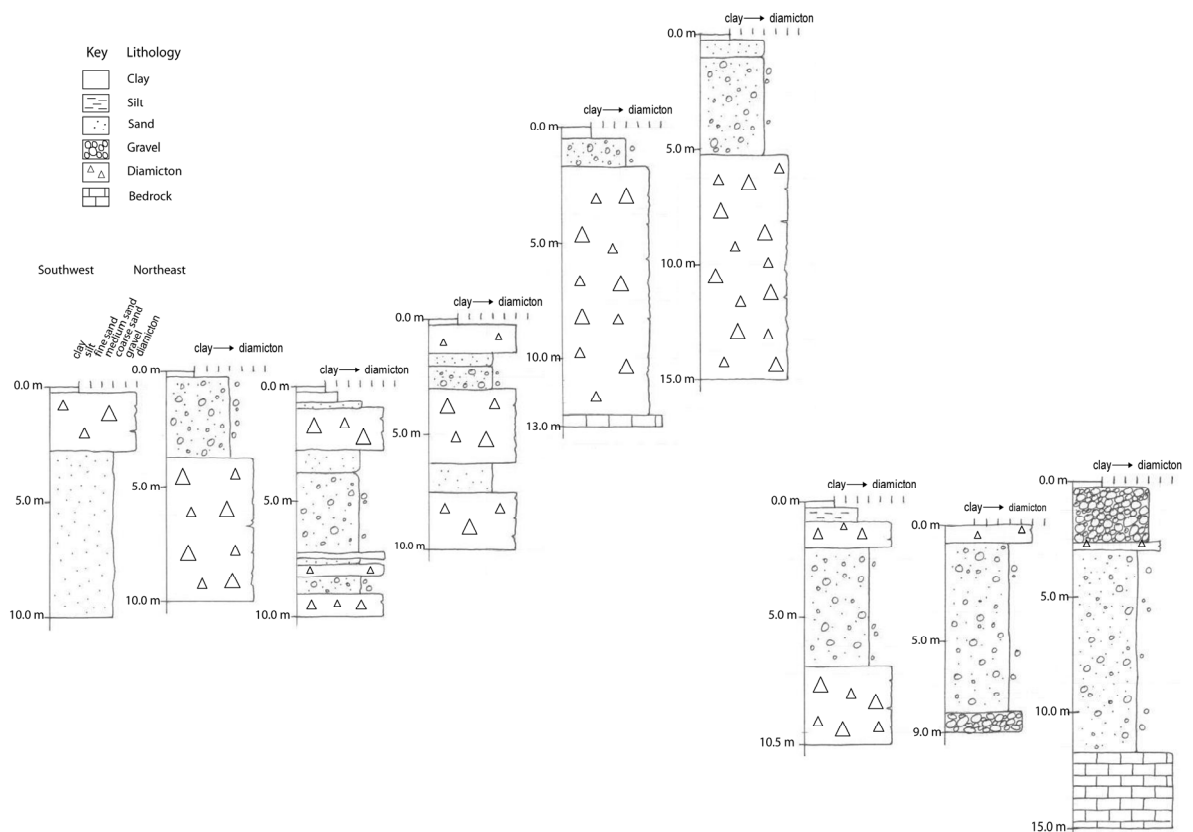


Figure 4.11: Borehole logs showing stratigraphic sections and borehole locations at Haltwhistle Bypass (BGS).

Hexham map sheet:

Mineral Assessment Report 65 (resource sheets NY 86, 96) for the Hexham region (Fig. 4.1c) of Northumberland describes a series of boreholes throughout the area (Lovell, 1981). The Hexham district lies between 30-60 m O. D. in the Tyne Valley and up to 300 m O. D. on the fells to the north and south. Located in the Carboniferous Namurian grits, the region contains predominantly west-east orientated flow phase 4 lineations. A series of borehole logs reveal 2 major lithofacies associations connected with the Late Devensian glaciation (Lovell, 1981).

Lithofacies association HX1 is a grey-brown diamicton resting on the Carboniferous bedrock. It ranges from a veneer to 60 m in thickness and contains a series of sand lenses. The diamicton is dominated by locally sourced rocks (sandstone, limestone and dolerite), with a minor component of Lake District (Borrowdale Volcanics and granite) and Scottish (granite) erratics (Lovell, 1981). This lithofacies also appears as thin layers locally overlying the gravel and sand of lithofacies association HX2.

Lithofacies association HX2 comprises coarse gravel and sand, with the gravel dominating in the Tyne Valley and sandier deposits towards the south-west (Lovell, 1981). These deposits form prominent hummocky features throughout the district. Lithologically HX2 is dominated by Carboniferous sandstone and Namurian limestone and also contains greywacke, Borrowdale Volcanics, quartzite and Scottish and Lake District granites (Lovell, 1981).

The diamicton is interpreted as a till by Lovell (1981). The foreign erratics offer some support for this interpretation, but taken in isolation there is insufficient evidence to be more specific and the diamicton can only be defined as glacial. The foreign erratics sourced from both Lake District and Southern Uplands indicate that the ice flows in the region produced a complex signature caused by overlapping trajectories and reworking of sediment. The thin upper diamicton has been interpreted as a solifluction or fluvial/lacustrine reworked deposit formed during the retreat of ice out of the Tyne Gap (Lovell, 1981). Lithofacies association HX2 is attributed to glaciofluvial activity due to its stratigraphic relationship with HX1, the presence of foreign erratics, the kamiform nature of deposition and the textural variability exhibited within it. The relatively fine-grained nature of the deposits suggests ice distal deposition.

Inter-site correlations:

Borehole information from Hexham map sheet, like Willowford, Hayden Bridge and Haltwhistle, displays a lowermost grey/brownish-black diamicton interpreted to be glacial. The colour, erratics and stratigraphic position of the glacial diamicton (lowermost unit, resting on bedrock) allows a tentative correlation to be made with other field sites, thus suggesting that deposition occurred over

a distance of at least 19 km. Indeed, previous research suggests that the lower grey diamicton is ubiquitous over most of Northern England (Huddart and Glasser, 2002; Clarke *et al.*, 2008). The two-tiered colour structure of the lower diamicton at Haltwhistle and the sand intercalations are also akin to the reddish and brownish-black diamicton sequence at Willowford (W4). This lateral continuity throughout the Tyne Gap region, together with the presence of Lake District and Scottish erratics means the diamicton is likely to be a subglacial till. Sands and gravels stratigraphically positioned above the lowermost diamicton and exhibiting high energy glaciofluvial conditions (cf. Miall, 1977) and upwards fining are pervasive throughout the Tyne Gap. The fining upward sequences within the sand and gravel lithofacies could relate to an increasingly ice-distal environment. The lack of diamicton within the deposits points towards a distal proglacial origin. The upper red diamicton is more disparate and generally thinner than the lower till where exposed. Genesis could be either as a re-advance till (Huddart and Glasser, 2002), or alternatively as a series of debris flows or re-worked deposits.

## 4.6 Discussion

### 4.6.1 *Ice sheet dynamics through the Tyne Gap corridor:*

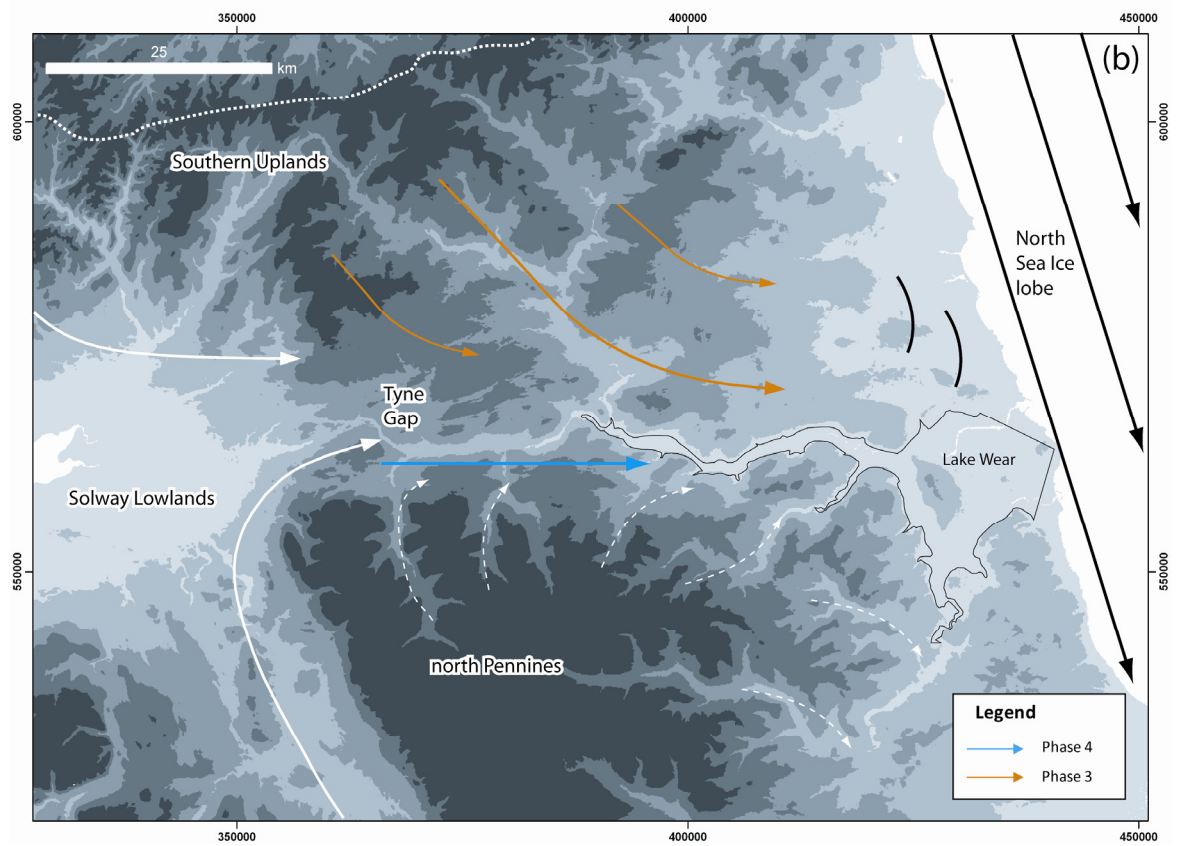
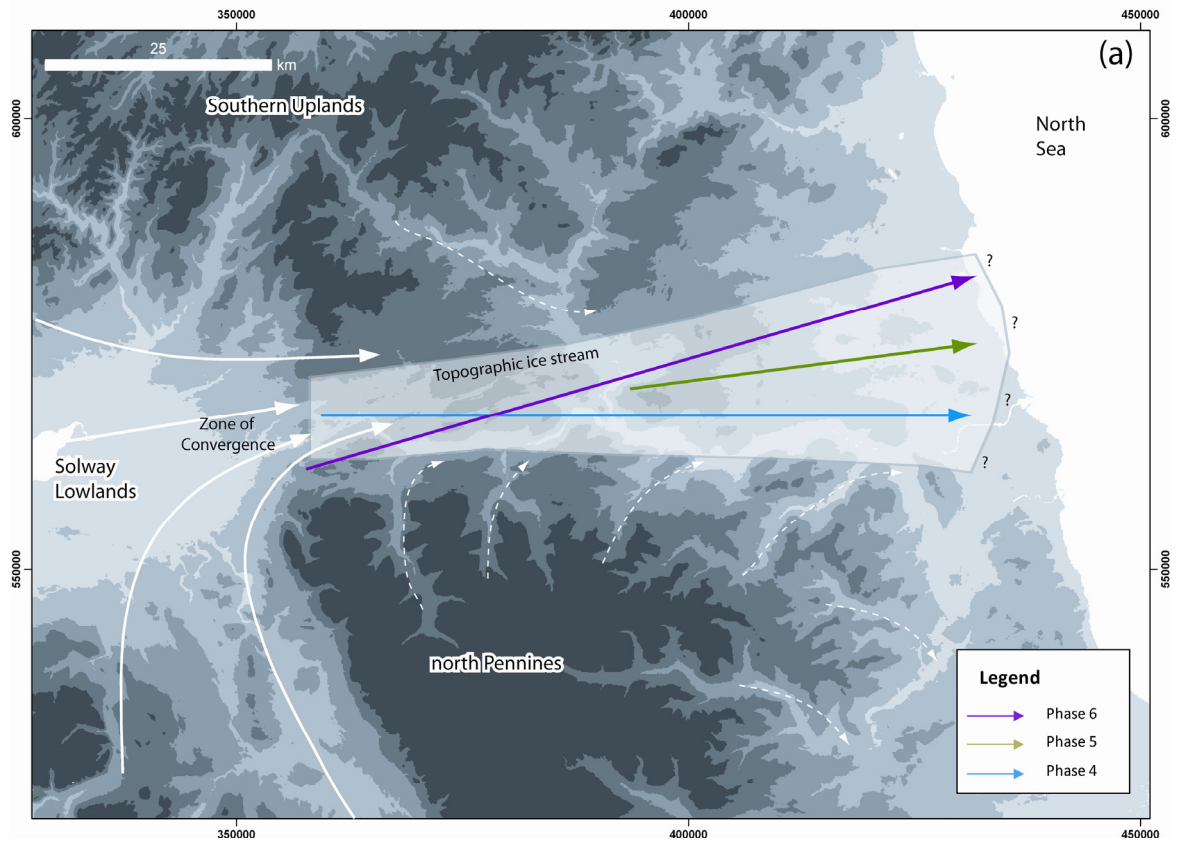
The ice flow phases and fans identified from glacial geomorphological mapping (Livingstone *et al.*, 2008) have been used to reconstruct three main stages of ice flow through the Tyne Gap. These stages are based on relative age of the ice flow phase, ice trajectory and subglacial conditions identified from the geomorphology and stratigraphy.

#### Stage I: The Tyne Gap as a major flow artery for converging Scottish and Lake District ice:

Flow through the Tyne Gap during stage I (Fig. 4.12a) was characterised by convergent ice flow sourced from the Scottish Southern Uplands and the Lake District, with locally sourced ice flowing into the main artery from the northern Pennines. Flow, as illustrated by the lineations and macro-fabric data from Willowford, converged into the Solway Lowlands, before overtopping the Tyne Gap col (152 m O. D.) and flowing to the east coast. This initial overtopping of the Tyne Gap followed a phase of glaciolacustrine sedimentation on the reverse slope at Willowford (see Fig. 4.4), with ice damming the pre-glacial fluvial system of the River Irthing. The diamicton on the reverse slope near Willowford was initially deposited in a series of debris flows (see Fig. 4.4), which reworked and incorporated the lacustrine sediments (cf. Eyles and Clague 1987; Eyles 1987; Bennett *et al.*, 2002).

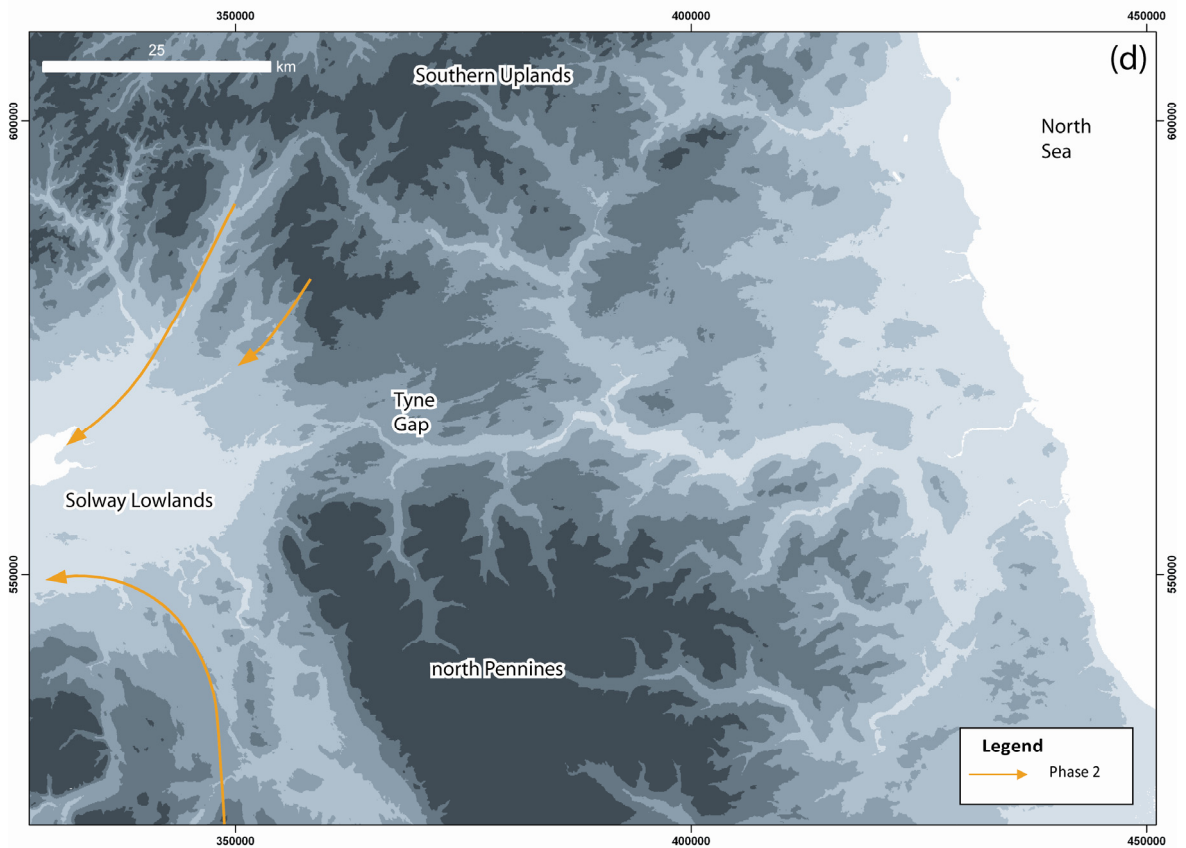
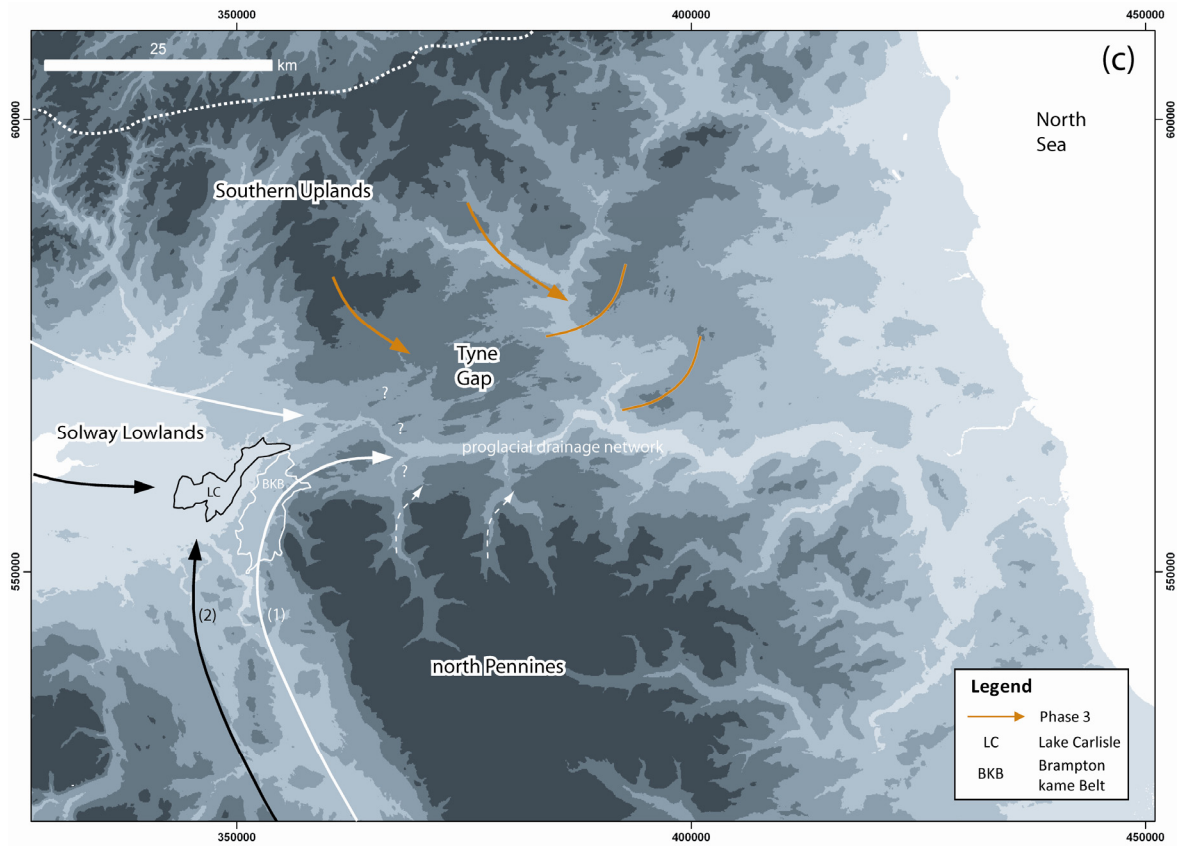


## CHAPTER 4: A MAJOR ICE DRAINAGE PATHWAY - THE TYNE GAP





## CHAPTER 4: A MAJOR ICE DRAINAGE PATHWAY - THE TYNE GAP



< Figure 4.12: Maps showing inferred ice flow stages within the Tyne Gap. Dotted lines indicate ice divides; dashed lines are streams of local ice movement: (a) Stage I: Flow through the Tyne Gap as a topographic ice stream (phases 6-4); (b) Stage II: Scottish ice breaching the eastward flowing ice artery, with south-easterly flow down the North Tyne. Initial retreat resulted in glacial Lake Wear forming between the North Sea ice lobe and ice stagnating within the Tyne Gap (phase 3); (c) Stage II/III: Westerly retreat of ice back into the Solway Lowlands resulting in a proglacial drainage network in the Tyne Gap, the formation of Brampton kame Belt and finally, the sequential development of glacial Lake Carlisle; and (d) Stage III: Drawdown of ice into the Irish Sea Ice Basin during the 'Blackhall Wood re-advance' (phase 2).

Geomorphological evidence suggests that ice flow through the Tyne Gap was initially dominated by Lake District ice, forcing flow in a north-easterly and then easterly direction (phase 6 and 5, Fig. 4.12a). Thus, during the earliest phases of flow recorded in the Tyne Gap, Scottish ice was subsidiary to that emerging from the Lake District. This is supported by the clast fabrics of the subglacial tills at Willowford (Fig. 4.7) which display a SW-NE orientation. This suggests that an ice divide became established in the northern sector of the Irish Sea Basin, south of the Southern Uplands, driving ice eastwards and causing coalescence with Lake District ice in the Solway Lowlands (Fig. 4.12a).

The corollary is that the earliest geomorphological signatures relate to phases of flow that occurred after southwards migration of the ice divide. This southward migration of the ice divide, observed at both the Firth of Clyde (Salt and Evans, 2004) and Isle of Man (Roberts *et al.*, 2007), was driven by ice expansion out of upland areas, particularly in Scotland (Boulton and Hagdorn, 2006). The indistinct provenance of the diamictons (Fig. 4.8) gives further credence to this argument, with a very complex build up of both Lake District and Scottish ice occurring within the Solway Lowlands (Hollingworth, 1931) and Tyne Gap, producing a mixed clast lithology (e.g. Willowford) due to cannibalisation and homogenisation of pre-existing diamictons.

During stage I the ice flow direction shifted from a SW-NE to a W-E orientation (Fig. 4.12a). This reflects a change in dominance between the contribution of Lake District and Scottish ice to the regional flow through the Tyne Gap and also the migration of the Solway Firth ice divide leading to re-establishment of Scottish flows across the region (flow phase 4). Shifts in ice-flow trajectory, which typify stage I, can therefore be reconciled by the migration of ice divides and ice dispersal centres in the surrounding source areas driving the changes in flow.

### Stage II: Scottish influenced ice flow down the North Tyne:

During stage II the growing influence of Scottish ice resulted in a major change of the ice flow in the Tyne Gap (Fig. 4.12b). The dominant movement in this stage was southeast, down the North Tyne Valley. This reflected continual northwards migration of the ice dispersal centre and

dissipation of the Solway Firth ice divide. As the ice divide migrated northwards, so the impact of the Tyne Gap as an easterly flow artery weakened. This resulted in the gradual recession and stagnation of the eastward margin of ice flow 5 leading to the time-transgressive deposition of kame deposits (Fig. 4.12c; Smythe, 1912), meltwater channels and transverse moraines at still-stand positions (Fig. 4.3). The geomorphological, sedimentological and stratigraphic evidence within the South Tyne Valley suggests that ice flow down the North Tyne Valley and out of Bewcastle Fells failed to reach the southern-most edge of the Tyne Gap. Geomorphological mapping also reveals similarly orientated south-easterly flow lineations sourced in Bewcastle Fells that continue into the Solway Lowlands and Tyne Gap (Fig. 4.3b). Thus, the ice divide must have straddled west-east across a large section of the Southern Uplands, allowing movement of ice into the Solway Lowlands and Tyne Gap along a broad front, with localised faster flow down palaeo-valleys (Fig. 4.12b). However, the elongation ratios suggest that these localised zones of faster flow were not comparable to the convergent flow during stage I (Table 4.1). Despite the weakened Lake District signal of phase II, a series of SW-NE eskers associated with the Brampton kame belt (which formed during stage III, Fig. 4.12c) indicate that ice continued to flow across the Tyne Gap at a late stage of glaciation (Fig. 4.12b). The aligned series of meltwater channels (on the flank of the northern Pennines) associated with phase 4 (Fig. 4.3d) are interpreted as an ice marginal drainage network, cut during a period of ice surface lowering towards the end of stage II and beginning of stage III when ice started to downwaste and retreat from the Tyne Gap.

#### Stage III: Drawdown of ice into the Irish Sea:

During the early part of stage III ice east of the Tyne Gap stagnated and retreated westwards (Fig. 4.12c) depositing a series of kamiform deposits within the Tyne valley (Yorke *et al.*, 2007). Borehole logs at Hexham and Haltwhistle and exposures at Hayden Bridge reveal a series of glaciofluvial sand and gravel deposits within the South Tyne Valley suggesting initiation of a major proglacial drainage system. The fining upwards sequence is typical of an increasingly ice-distal environment, with the discontinuous, thin upper diamicton thought to relate to either debris flow deposition or a minor ice-marginal oscillation. As ice retreated onto the reverse slope at Brampton, ice stagnated in the lee of the Pennines (Fig. 4.12c).

Following the dissipation of the Solway Firth ice divide and the partial deglaciation of the Solway Lowlands, ice subsequently re-advanced ('Blackhall Wood re-advance') into the regional topographic low of the Irish Sea Basin (Livingstone *et al.*, in press) (Fig. 4.12d). The rapid streaming of the Irish Sea ice lobe (Eyles and McCabe, 1989; Evans and Ó Cofaigh, 2003; Ó Cofaigh and Evans, 2007; Thomas and Chiverrell, 2007) facilitated drawdown in the Solway region. This had a major effect on ice flow trajectories, as Scottish ice from the central Southern Uplands (associated with the maintenance of an ice divide in this region) switching direction,

flowing SW across the Tyne Gap (out of Bewcastle Fells), and into the Solway Lowlands (phase 2; Fig. 4.12d).

The progressive retreat of ice into the Solway Lowlands followed by ice flow out into the Irish Sea ice basin ended the eastwards transfer of ice and resulted in the Tyne Gap being cut-off from ice flow. Subsequent flows (phase 1) were topographically constrained with no evidence for encroachment into the Tyne Gap (Salt and Evans, 2004; Livingstone *et al.*, 2008).

#### 4.6.2 *A palaeo-ice stream?*

The regional ice flow patterns during stage I display several features consistent with the former presence of an ice stream in the Tyne Gap. Flow from the Lake District and Southern Uplands converged on the Tyne Gap, to the east of the Solway Lowlands, before entering a narrow, 25 km wide, topographically-constrained trunk between Cold Fell to the south and the Bewcastle Fells to the north. Flow then followed an easterly trajectory through the Tyne Gap, represented by a coherent, 50 km long, W-E orientated flow set of lineations (Figs. 4.3 and 4.12). The true length of the ice flow trunk is not known as the lineations have been overprinted by N-S orientated bedforms, resulting in the gradual disappearance of the W-E orientated forms towards the east coast. It is not known whether the Tyne Gap ice was confluent with ice moving south from the Cheviots and Tweed valley during stage I or if it formed a lobate margin in the North Sea Basin. Throughout stage I, a series of ice flow shifts occurred, suggesting that the Tyne Gap ice stream was not static. These dynamic shifts could relate to a time transgressive ice stream imprint (Clark, 1999). This transient behaviour has been observed in modern examples, such as ice stream B (Bindshadler and Vornberger, 1998) and C (Anandakrishnan and Alley, 1997) in Antarctica.

Although stage I lineations are highly attenuated and well developed, their elongation ratios are only between 2:1 and 4:1. These are significantly lower than those documented from other palaeo-ice streams (e.g. Stokes and Clark, 1999; 2001). Furthermore no mega-scale glacial lineations have been observed in the Tyne Gap. It is possible that this reflects the influence of substrate control in that the lineations that do occur are predominantly rock moulded. The dynamic flow switches observed during stage I also suggest that ice flow may have been transient. Therefore, of the three main factors observed to be accountable for attenuated bedforms in drumlin fields (cf. Stokes and Clark, 2002), the ice flow duration and sediment supply in the Tyne Gap are not conducive to the formation of highly elongate features, thereby making it difficult to quantify the relative ice flow velocities.

The stratigraphic and sedimentological data from Willowford suggests that subglacial deformation did occur in the Tyne Gap (Fig. 4.4). Although homogenisation is not complete, the clast macrofabric data (Fig. 4.7) throughout LFA W4, coupled with intercalated tills,

micromorphological evidence of rotational and planar structures, and rafted and crushed bedrock all suggest that at Willowford subglacial stresses were sufficient to deform the substrate. However, although evidence of till deposition within the Tyne Gap has generally been recorded in the tripartite division of grey till - sands and clays - red till (Hughes *et al.*, 1998), glacial diamicton (till) is patchy throughout the region and large areas are characterised by streamlined bedrock. This widespread sparsity of till does not predispose against streaming, as ice streams over hard beds have been proposed (e.g. I.S. Evans 1996; Stokes & Clark 2003; Roberts and Long, 2005; Bradwell *et al.*, 2008). It does, however, suggest that soft-bedded flow is unlikely to be a significant mechanism for streaming in the Tyne Gap.

The discussion above suggests that there is evidence for a topographic ice stream in the Tyne Gap. However, relative flow velocities are not significantly different to other phases of flow within the region, suggesting that any streaming that did occur was topographically induced. Indeed the elongation ratios are not highly attenuated ( $>10:1$ ) although the general mosaic of bedrock and drift deposits coupled with the dynamic shifts in ice flow offers some explanation for this. Thus the evidence that the Tyne Gap did act as a major artery for flow transferring a disproportionate flux of ice out of the Solway Lowlands from both Scottish and Lake District sources is compelling.

#### 4.6.3 *Wider implications for the British-Irish Ice Sheet*

During the Late Devensian it has been proposed that a tripartite glacial sequence was deposited along the east coast of England. First, the 'lower boulder clay' (Smith, 1981) is a greyish-blue till containing abundant Carboniferous rocks, deposited by easterly moving ice (Raistrick, 1931; Catt, 1991). This till correlates with the eastward movement of ice through major arteries in the Pennines (Douglas, 1991) including the Tyne Gap. Consequently, the lower till can be stratigraphically correlated to stages I and II of flow through the Tyne Gap. Second, overlying the lower till is a series of 'middle sands', interpreted by Smith (1981) as the glaciolacustrine deposits of Lake Wear. These deposits are understood to have been deposited between westerly retreating ice and a southwards flowing Tweed-Cheviot ice stream off the east coast (Teasdale and Hughes, 1999). This equates to flow stage II/III involving Tyne Gap stagnation and westerly retreat (Fig. 4.11c). The Tweed-Cheviot ice stream flowing southwards down the east coast was thought to be in existence during flow stage I and II, although the Tyne Gap was dominant (Raistrick, 1931). Third, during the later stages of glaciation (associated with flow stage III) the Tweed-Cheviot ice became more dominant, pushing inland (cf. Catt, 1991) and depositing the upper red diamicton (Smythe, 1912; Davies *et al.*, 2009).

Thus the stratigraphy of the east coast of England around Northumbria and County Durham can be reconciled with the geomorphological and stratigraphic evidence within the Tyne Gap. The Tyne

Gap ice initially dominated ice dynamics in south Northumbria and County Durham, impeding or deflecting the Tweed-Cheviot ice stream further eastwards. This was followed by a more dominant Cheviot-Tweed ice flow (Davies *et al.*, 2009). Runoff from the downwasting Tyne Gap ice is the most likely source of the water for the expansion of Lake Wear before the Tweed-Cheviot ice advanced westwards, re-organising the bedform pattern and partially erasing the evidence of the Tyne Gap ice stream.

It is concluded here that south-west flow into the Irish Sea Ice Basin (Livingstone *et al.*, 2008) only occurred during ice flow stage III. This late stage re-advance and switch in ice flow direction is thought to be associated with the dissipation of the Solway Firth ice divide and the migration of the ice dispersal centre northwards back into the Southern Uplands. With flow no longer being forced eastwards through the Tyne Gap, ice was free to drain into the topographic low of the Irish Sea Basin.

During the last glaciation the Irish Sea Ice Basin was occupied by an ice stream (e.g. Evans and Ó Cofaigh, 2003; Ó Cofaigh and Evans, 2007) which reached as far south as the Scilly Islands (Scourse, 1991; Hiemstra *et al.*, 2005) during the Last Glacial Maximum. On the Isle of Man, in the Irish Sea Basin, two phases of flow have been identified (Roberts *et al.*, 2007), with phase II interpreted as a late stage advance initiated in the Solway Lowlands. Thus, the relative chronology of glacial events in the Tyne Gap ties in with drawdown of ice into the Irish Sea Ice Stream. Initial flow of ice during flow stages I and II, relates to phase I on the Isle of Man (Roberts *et al.*, 2007), with the western Southern Upland and Highland ice providing most of the ice flux. This is consistent with flow phases B-C identified by Salt (2001) and Salt and Evans (2004) in south-west Scotland; Highland ice from the Firth of Clyde coalesced with Southern Upland ice forming a powerful south-southwest flow into the Irish Sea. This was postulated to have occurred during the Last Glacial Maximum, when the Irish Sea Ice Stream was at its maximum (phase C; Salt, 2001, Salt & Evans 2004). These powerful southwards moving ice flows buttressed and compressed the Lake District and Solway Lowland ice to the west, forcing it to overtop and flow eastwards through the Tyne Gap. During deglaciation Tyne Gap and Solway Lowland ice acted as a major ice source for the Irish Sea Ice Stream (phase II; Roberts *et al.*, 2007). In the south-west Southern Uplands the ice divide migrated eastwards during stage D-F (Salt, 2001), causing the central Southern Uplands to become increasingly important, and thereby supplying the majority of ice down the North Tyne (stage II) and eventually allowing ice to drain south-west into the Irish Sea (stage III) as the influence of Highland and south-western Southern Upland ice began to wane. This allowed the Tyne Gap to act as part of a major Solway Lowlands ice stream tributary, as ice was drawn down into the Irish Sea Basin. Eventually topographically constrained flow became pervasive throughout the area (phase F, Salt, 2001; phase 1, this paper), with independent glaciers occupying valleys.



## 4.7 Conclusions

- Flow sets of streamlined subglacial bedforms provide evidence for former dynamic flow within the Tyne Gap during the last glaciation.
- On the basis of the glacial geomorphological mapping 3 main stages of ice flow in the Tyne Gap are recognised.
  - ⇒ Stage I is associated with major arterial ice flow moving generally eastwards driven by ice sourced from both Scotland and the Lake District. Evidence for dynamic shifts in flow throughout this stage are indicative of migratory ice divides reflecting a change in dominance between Scottish and Lake District dispersal centres.
  - ⇒ Stage II was dominated by Scottish ice moving south-easterly down the North Tyne Valley and deflecting the eastwards flowing ice artery. This reflects the continual northwards migration of the ice dispersal centre and dissipation of the Solway Firth ice divide.
  - ⇒ During stage III the influence of ice within the Tyne Gap waned, resulting in progressive deglaciation westwards into the Solway Lowlands. During the ‘Blackhall Wood re-advance’ ice drained SW into the Irish Sea Ice Basin as a tributary of the Irish Sea Ice Stream.
- For at least part of this history (stage I) flow through the Tyne Gap is inferred to have been in the form of a topographic ice stream. Key indicators include a strongly convergent geometry, streamlined subglacial bedforms and some evidence for deformable bed conditions.
- During deglaciation a major proglacial drainage network developed in the Tyne Valley as ice retreated progressively westwards. As ice retreated across the Tyne Gap watershed ice stagnated against the reverse slope, leading to the formation of Lake Carlisle and the Brampton kame belt.

## 4.8 Acknowledgements

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## Chapter 5

# **The Brampton kame belt and Pennine escarpment meltwater channel system (Cumbria, UK): Morphology, sedimentology and formation**

Livingstone, S.J., Ó Cofaigh, C. & Evans, D.J.A., Hopkins, J. In press. The Brampton Kame Belt and Pennine Escarpment meltwater channel system (Cumbria, UK): Morphology, Sedimentology and Formation. *Proceedings of the Geologists Association*.

## **Abstract**

The Brampton kame belt represents one the largest glaciofluvial complexes within the UK. It is composed of an array of landform and sediment assemblages, associated with a suite of meltwater channels and situated within a palimpsest landscape of glacial features in the heart of one of the most dynamic parts of the British-Irish Ice Sheet. Glacial geomorphological mapping and sedimentological analysis has allowed a detailed reconstruction of both the morphological features and the temporal evolution of the Brampton kame belt, with processes informed by analogues from modern ice margins. The kame belt demonstrates the development of a complex glacier karst typified by the evolution of subglacial meltwater tunnels into an englacial and supraglacial meltwater system dominated by ice-walled lakes and migrating ice-contact drainage networks. Topographic inversion led to the extensive re-working of sediments, with vertical collapse (associated with the melt-out of dead ice) and debris flows causing partial disintegration of the morphology. The resultant landform comprises a series of kettle holes, discontinuous ridges (eskers) and flat-topped hills (ice-walled lake plains). The Pennine escarpment meltwater network, which fed the Brampton kame belt, is composed of an anastomosing subglacial channel system and flights of lateral channels. The Brampton kame belt is envisaged to have formed during the stagnation of ice in the lee of the Pennines and Penrith sandstone ridge as ice retreated westwards across the Tyne Gap into the Solway Lowlands, prior too, during and following the 'Blackhall Wood re-advance'. The formation of the Brampton kame belt also has particular conceptual resonance in terms of constraining the nature of kame genesis, whereby an evolving glacier karst is a key mechanism in the spatial and temporal development of ice-contact sediment-landform associations.

## 5.1 Introduction

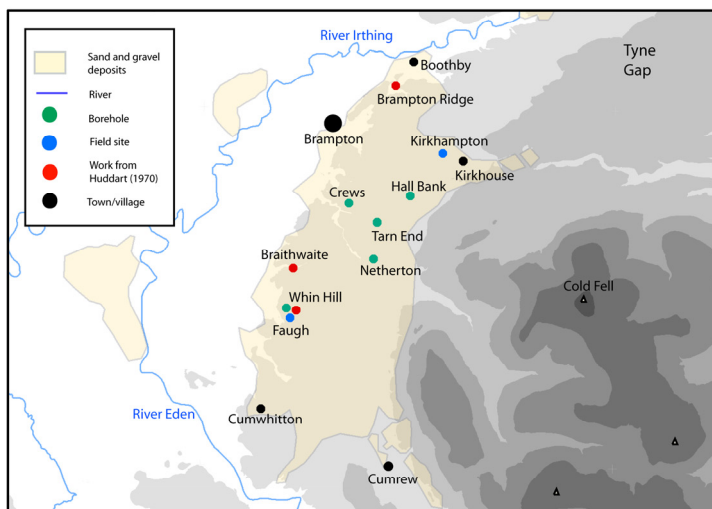
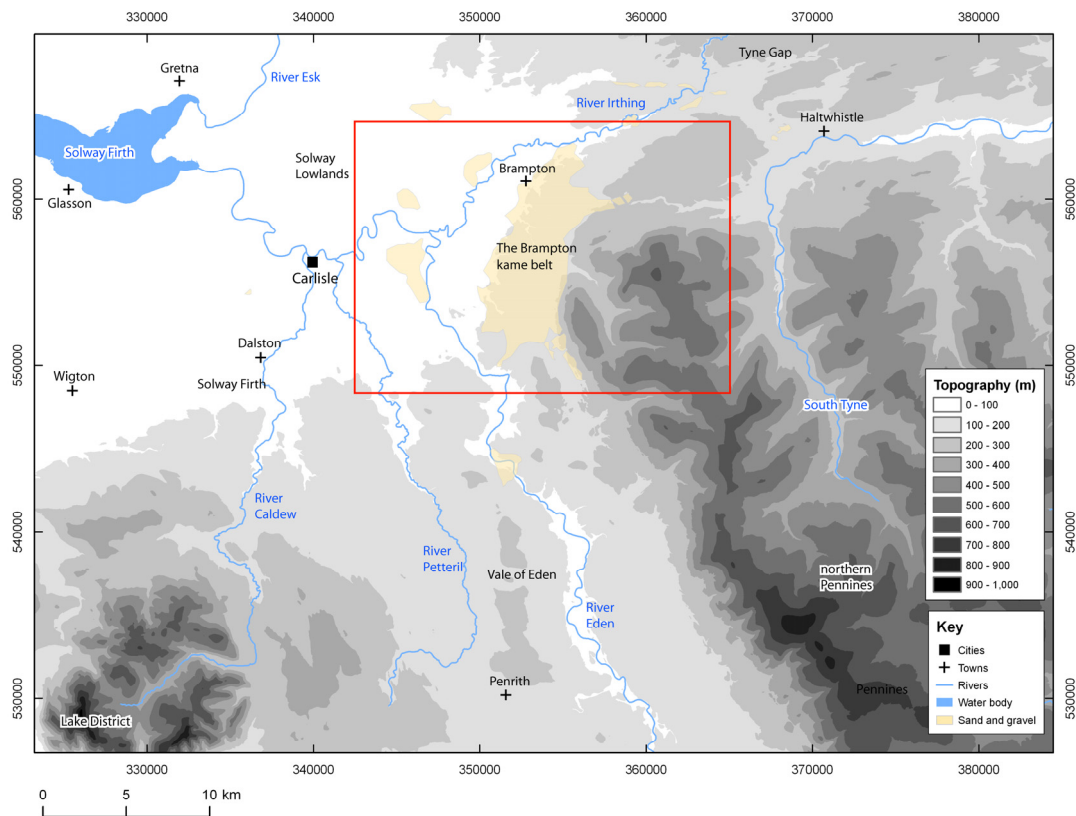
Kames constitute some of the most diverse and complex landform and sediment assemblages in glaciated landscapes. Their morphology often comprises an array of ridges, flat-topped hills and depressions which contain a diverse range of glaciofluvial and glaciolacustrine sediments (e.g. Holmes, 1947; Price, 1969, 1973; Paul, 1983; Thomas *et al.*, 1985; Benn & Evans, 1998; Evans & Twigg, 2002; Evans *et al.*, 2009a). Consequently the site-specific interpretation of these landforms is often contentious (cf. Gravenor & Kupsch, 1959; Huddart & Bennett, 1997; Owen, 1997; Thomas & Montague, 1997; Johnson & Clayton, 2003). However, observations of glaciofluvial landform-sediment assemblages along modern glacier margins (e.g. Price, 1966, 1969, 1973; Huddart, *et al.*, 1999; Russell, *et al.*, 2001, 2005, 2007; Evans & Twigg, 2002; Evans *et al.*, 2009a) are proving increasingly useful in constraining the genesis of kames and their associations with ice-contact fans and esker networks. In light of recent advances it is appropriate to re-appraise models of kame formation in Pleistocene environments. The “Brampton kame belt” in Cumbria (cf. Trotter, 1929; Trotter & Hollingworth, 1932; Huddart, 1970, 1981) represents one of the largest assemblages of glaciofluvial material in the United Kingdom at over 44 km<sup>2</sup> (Livingstone *et al.*, 2008). It represents a major depositional episode during the advanced stages of recession of the Late Devensian (Dimlington stadial) British and Irish Ice Sheet (BIIS) in the Solway Lowlands (Trotter, 1929; Huddart, 1970, 1981). Being able to elucidate the mode of formation of such a large feature, together with its genetic association with a complex suite of meltwater channels on the Pennine escarpment (Fig. 5.1) to the south (Arthurton & Wadge 1981), is critical to reconstructing the style of deglaciation at the centre of the BIIS.

This paper has two objectives: to determine the origin of the various components of the Brampton kame belt; and to use this information to constrain the nature, configuration and timing of deglaciation of the area comprising the Solway Lowlands, the Vale of Eden/Pennine escarpment and westernmost Tyne Gap, the core region at the centre of one of the most dynamic parts of the BIIS (Salt & Evans 2004; Livingstone *et al.*, 2008; Evans *et al.*, 2009b). Resolving these objectives will enable the development of a depositional model for the formation of the Brampton kame belt, which in turn will allow general conclusions regarding kame genesis to be elucidated.

## 5.2 Process-form models of kame production and previous work on the Brampton area

The term “kame” is used by glacial geomorphologists to refer to “steep-sided, variously shaped mounds composed chiefly of sand and gravel, formed by supra-glacial or ice-contact glaciofluvial deposition” (Benn & Evans, 1998, p487). Kame topography forms a composite landscape consisting of flat-topped mounds, ridges and hollows (Fig. 5.2) that evolve when large quantities of

sediment are reworked by englacial and supraglacial drainage networks during the final retreat of ice (Cook, 1946; Holmes 1947; Paul 1983). In general terms hollows are formed by the differential melting of debris-covered dead ice and the enlargement and collapse of tunnels during downwasting (Clayton, 1964; Clayton & Cherry, 1967; Krüger, 1994). The deposition of glaciofluvial, glaciolacustrine and interfingering debris flow deposits in the hollows initiates continuous topographic inversion during ablation, forming positive relief features (Boulton, 1967, 1972; Price, 1965, 1969). The removal of supporting ice produces complex patterns of fault and fold structures within the kame deposits (Boulton 1972; McDonald & Shilts, 1975; Johnson & Clayton, 2003).





< Figure 5.1: Topographic map showing the position of the Brampton kame belt within the central sector of the BIIS, and the location of fieldsites and borehole logs.

Flat-topped mounds (kame plateaus) are the product of deposition in ice-walled lake plains (e.g. Smed 1962; Clayton & Cherry, 1967; Johnson & Clayton, 2003; Clayton *et al.*, 2008) or supraglacial ponds (e.g. Mager & Fitzsimons, 2007). Supraglacial ponds collect sequences of heavily disturbed stratified sediments and mass flow deposits which typically show evidence of water escape structures, soft sediment deposition, and cut and fill sequences (Eyles, 1979; Eyles *et al.*, 1987; Mager & Fitzsimons, 2007). Ice-walled lake plains are characterised by flat-tops, rhythmically bedded fine-grained sediments and deltaic deposits. At the margins of the upstanding lake plains or plateaux these sediments are typically extensively faulted due to the removal of the supporting ice walls during melt-out (cf. Clayton *et al.*, 2008).

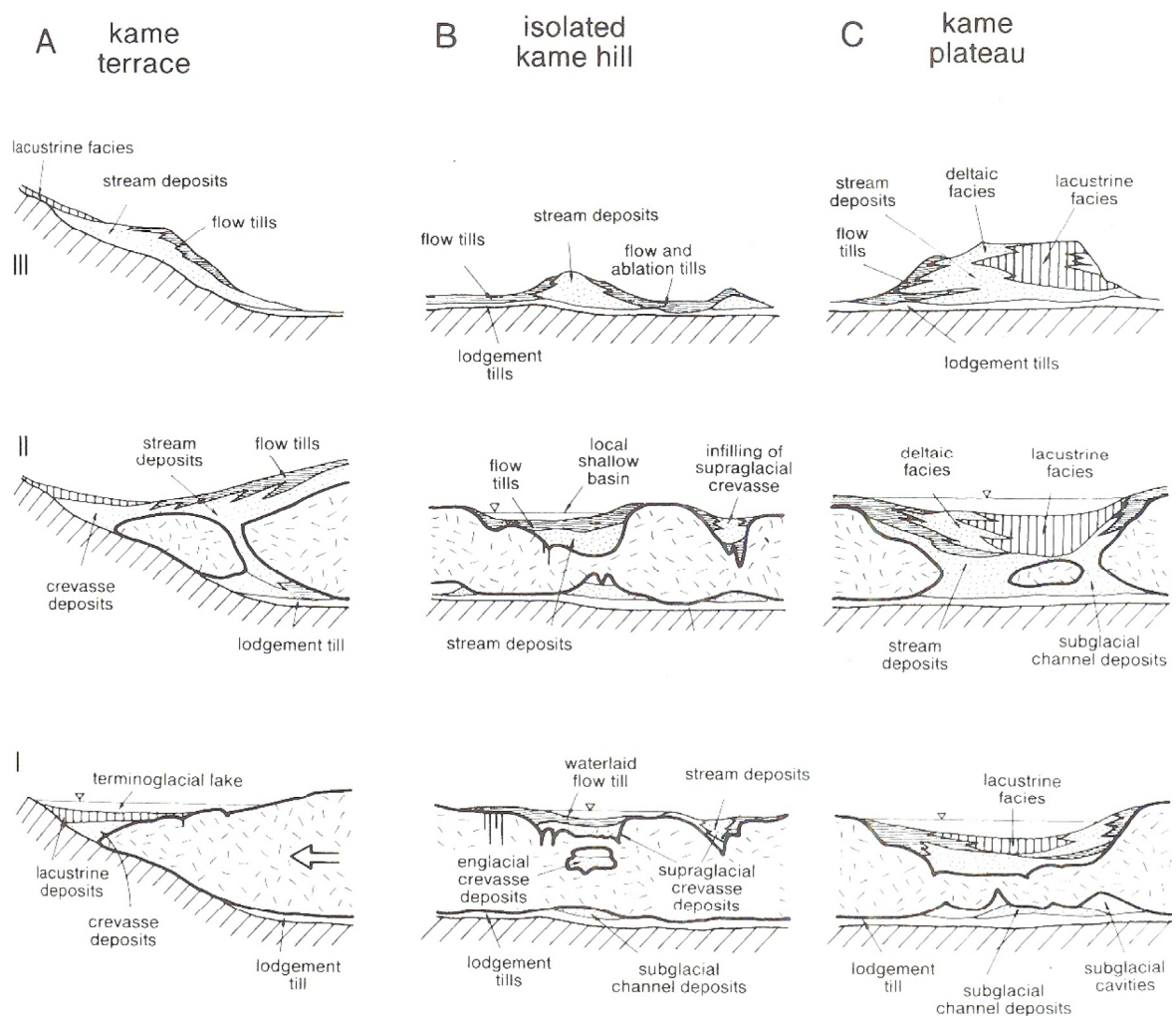


Figure 5.2: Schematic model for the various types of kames (from Brodzikowski & van Loon, 1991).

Kame terraces are “gently sloping depositional terraces perched on valley sides...deposited by meltwater streams flowing between glacier margins and the adjacent valley wall” (Benn & Evans, 1998, *p.* 490). Thus a series of kame terraces can document the periodic retreat and surface lowering of a glacier (Sissons, 1958). Kame terraces typically exhibit pitted surfaces due to the

melt out of buried ice. Substantial bodies of glacier ice may lie beneath the deposits of marginal streams (McKenzie 1969; Evans *et al.* 2009a) making it unlikely that the flat river bed will survive final melt-out, and will develop instead into an elongate chain of kame and kettle topography in which the kame mound summits appear accordant from a distance.

Kame topography often contains features whose form is dictated by the palaeo-drainage network or guidance of supraglacial streams by controlled moraine (*sensu* Evans, 2009) produced by debris-charged folia rising to the glacier surface (e.g. Thomas *et al.*, 1985). Consequently, discontinuous sinuous ridges composed of glaciofluvial sediments are visible amongst the kame and kettle terrain (e.g. Price, 1969; Twigg & Evans, 2002). These esker-like ridges often show a great deal of complexity (e.g. the “Carstairs kames” in Scotland; Huddart & Bennett, 1997; Thomas & Montague, 1997), necessitating multiple phases or mechanisms of construction. Glaciofluvial sediment can accumulate at various levels in a glacier and may evolve from a subglacial to an englacial and supraglacial assemblage during ice wastage (e.g. Huddart, *et al.*, 1999).

The Brampton kame belt was first investigated by Trotter (1929) (Fig. 5.3). He proposed a model of ice frontal retreat, with ridges attributed to ice-contact outwash formed during stillstands. These were fed by eskers [“oses” in Trotter, 1929] which were interpreted as englacial conduits (Fig. 5.3). These complex esker deposits, often consisting of anastomosing limbs, were attributed to the crevassed condition of the ice which produced multiple changes in stream course. Flat-topped hills marked ice-marginal delta fronts (Fig. 5.3), again fed from englacial streams, built up to lake level. The circular hollows were interpreted by Trotter as kettle holes formed by ablating dead ice. The kame belt was re-investigated by Huddart (1970) who attributed the landforms to supraglacial deposition during *in situ* ice stagnation, rather than the ice-marginal depo-centres envisaged by Trotter. The depositional ridges in the kame belt were interpreted by Huddart as the products of a rapidly shifting glaciofluvial drainage network following open crevasses. The network was fed from englacial or subglacial tunnels. The flat-topped hills were interpreted as standing bodies of water surrounded by stagnant ice (i.e. ice-walled lake plains).

The Brampton kame belt is contiguous with a series of meltwater channels and glaciofluvial deposits that extend southwards along the Pennine escarpment and northeastwards into the Tyne Gap (Figs. 5.3 and 5.4) (Trotter, 1929; Arthurton & Wadge, 1981; Livingstone *et al.*, 2008). The Hallbankgate esker, which emerges from the northern end of the kame belt, trends into an 8 km long meltwater channel cutting west to east across the Irthing-South Tyne watershed (Fig. 5.3) and that feeds into the Tyne Valley system (Trotter, 1929; Livingstone *et al.*, 2008). Another channel to the north of the Brampton kame belt at Gilsland cuts NW-SE across the Irthing-South Tyne watershed (Fig. 5.3). This is associated with a branching complex of glaciofluvial deposits, with one arm trending SW towards the Brampton kame belt and one arm trending NW towards Scotland (Trotter, 1929; Fig. 5.3). Feeding into the Brampton kame belt from the south is a network of SE-

NW orientated meltwater channels cutting obliquely downslope in a northward direction along the Pennine escarpment (Figs. 5.3 and 5.4) (Trotter, 1929; Arthurton & Wadge, 1981; Clark *et al.*, 2004; Evans *et al.*, 2005; Greenwood *et al.*, 2007; Livingstone *et al.*, 2008). These have been previously interpreted either as a series of aligned marginal channels (Trotter, 1929) or as a predominantly subglacial drainage network with lateral and ice marginal feeder channels higher up on the escarpment flanks (Arthurton & Wadge, 1981; Greenwood *et al.*, 2007).

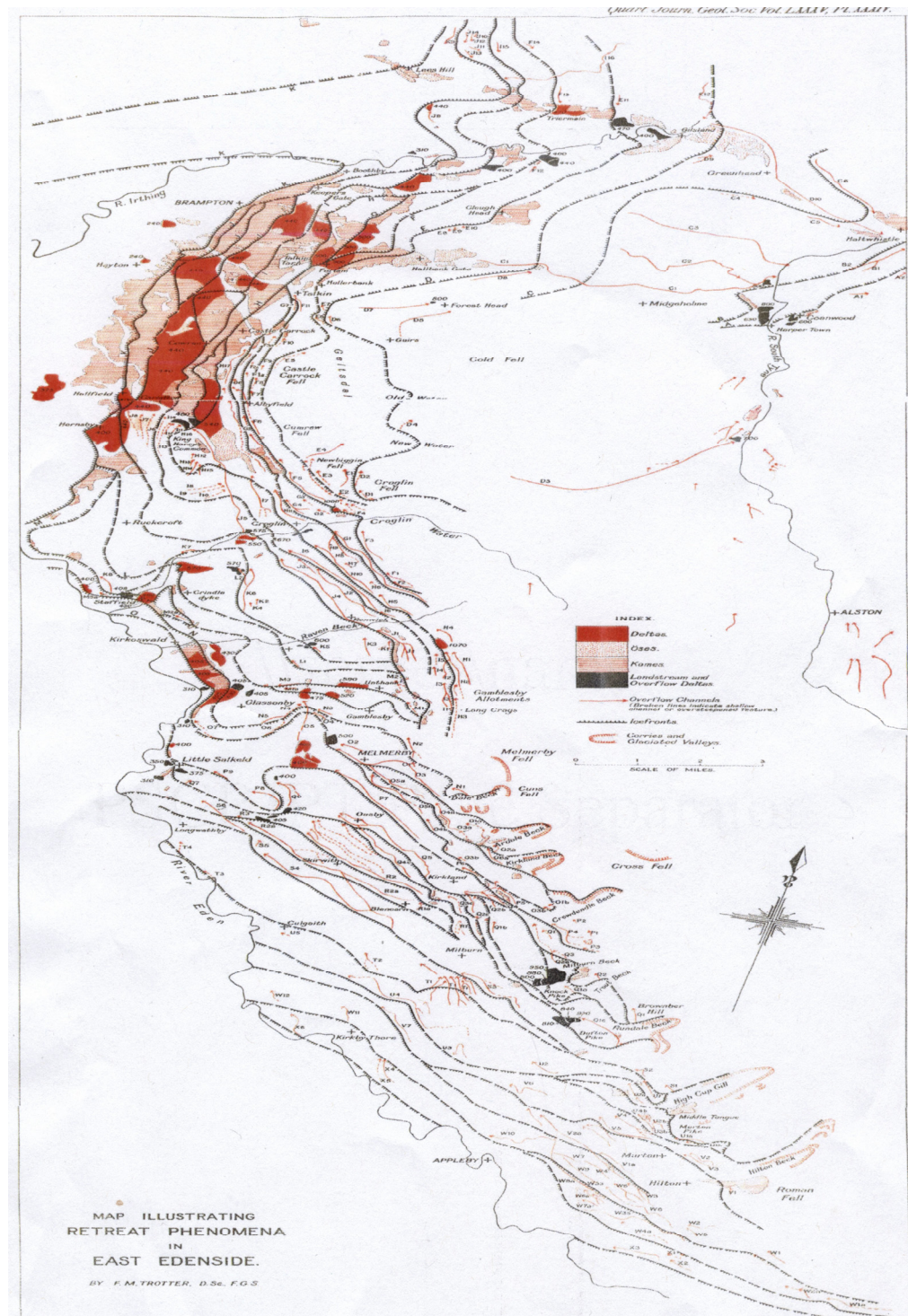


Figure 5.3: Retreat phenomena in east Edenside demarcating a series of ice marginal positions from the kame belt and meltwater channels (from Trotter, 1929).



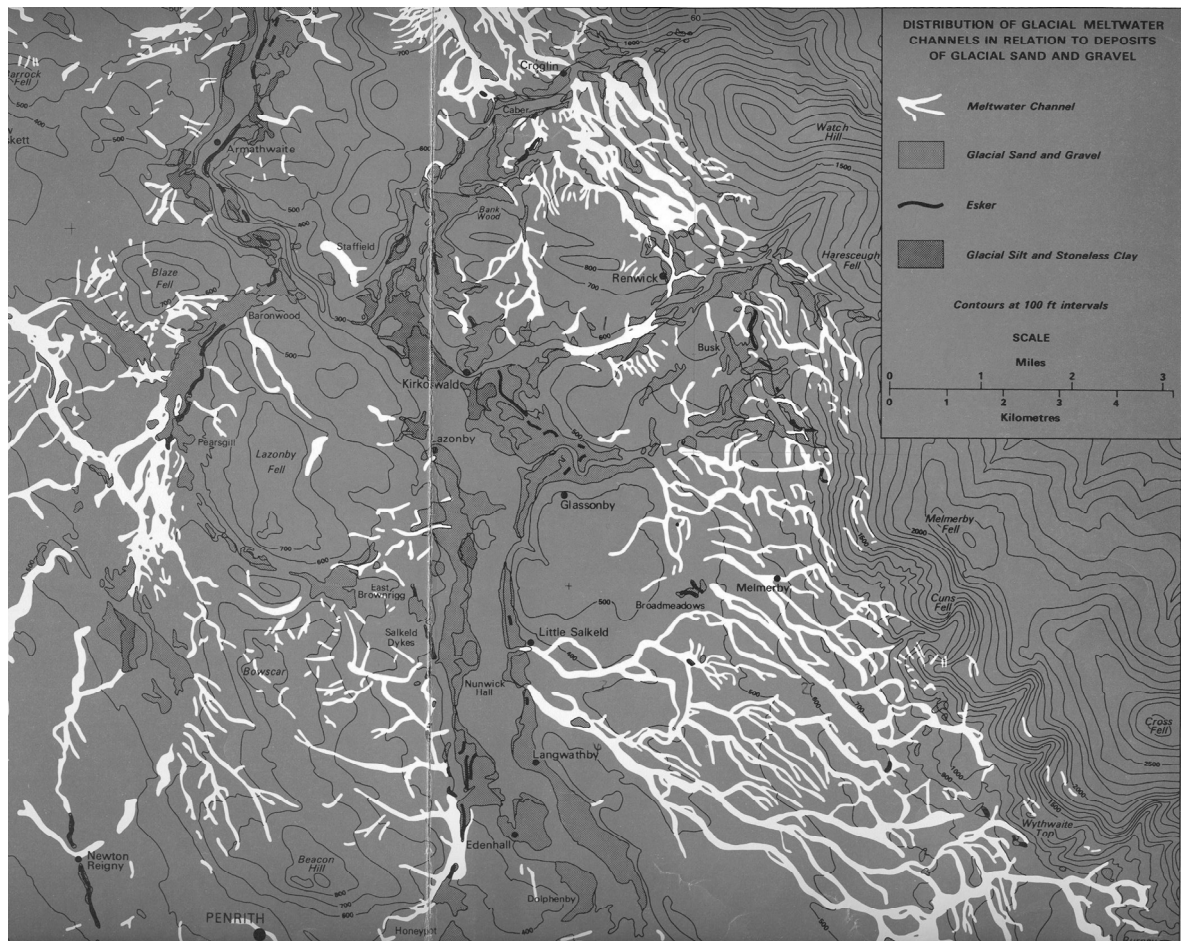


Figure 5.4: Meltwater channels along the Pennine escarpment and within the Penrith sandstone ridge (from Arthurton & Wadge, 1981).

### 5.3 Methods

#### 5.3.1 Geomorphological mapping

Geomorphological mapping of the Brampton kame belt and surrounding area involved the recognition and compilation of discrete landform assemblages both on NEXTMap 5 m resolution airborne Interferometric Synthetic Aperture Radar (IFSAR) imagery (<http://www.neodc.rl.ac.uk/>), and digitised, geo-rectified aerial photographs (Cambridge University). The NEXTMap data was organised into National Grid tiles of 10 x 10 km based on Ordnance Survey co-ordinates. These data have been acquired by the British Geological Survey (BGS) for NERC and are archived at the NERC Earth Observation Data Centre (NEODC). Vertical aerial photographs allowed mapping at a very high resolution, thus allowing further features to be picked out that were not visible on the NEXTMap DEM. Landforms were identified, based purely on morphological characteristics, into ridges (vectors), meltwater channels (vectors) depressions (polygons) and flat-topped hills (polygons). Mapping was carried out manually using on-screen digitisation and ground truthing used to verify specific features. Layers were draped over the NEXTMap and aerial photograph

imagery to: (a) display topographic contours (created from the NEXTMap DSME); (b) remove anomalous human artefacts (Ordnance Survey); (c) reveal the digitised results of the Glacial Map of Britain (Clark *et al.*, 2004; Evans *et al.*, 2005); and (d) illustrate both the superficial sedimentary deposits and bedrock geology (BGS) within the field area. Profiling of the meltwater channels along the Pennine escarpment was carried out digitally by generating nodes at 5 m intervals along the meltwater channel vectors and assigning each vector an x, y and z co-ordinate.

### 5.3.2 *Sedimentology and stratigraphy*

Borehole information and exposures in both ridges and flat topped hills at Kirkhouse and Faugh sand quarries in the Brampton kame belt have been used to interpret depositional environments for the various geomorphological features. Texture, sedimentary structure, colour (Munsell colour chart), bed geometry, contacts and inclusions were all measured and logged, from which lithofacies associations were identified. Lithofacies codes are based upon those of Evans and Benn (2004). Scaled section sketches were drawn at the larger exposures so that the lateral extent of the lithofacies could be assessed. Paleaocurrent indicators (such as ripples and imbricate gravel), and clast lithologies augmented the sedimentary logging. The sampling procedure for clast lithological analysis is based on those outlined in Bridgland (1986) and Walden (2004), with three size ranges adopted, 8-16 mm, 16-32 mm and 32-64 mm, and a sample size of  $\geq 300$ . Identification of lithologies was carried out using a hand lens and an optical microscope. Additional sedimentological and stratigraphic data from exposures in other sand quarries was obtained from Huddart (1970). Borehole logs (Jackson, 1979) provided a less detailed but wider coverage, allowing regional stratigraphic correlations.

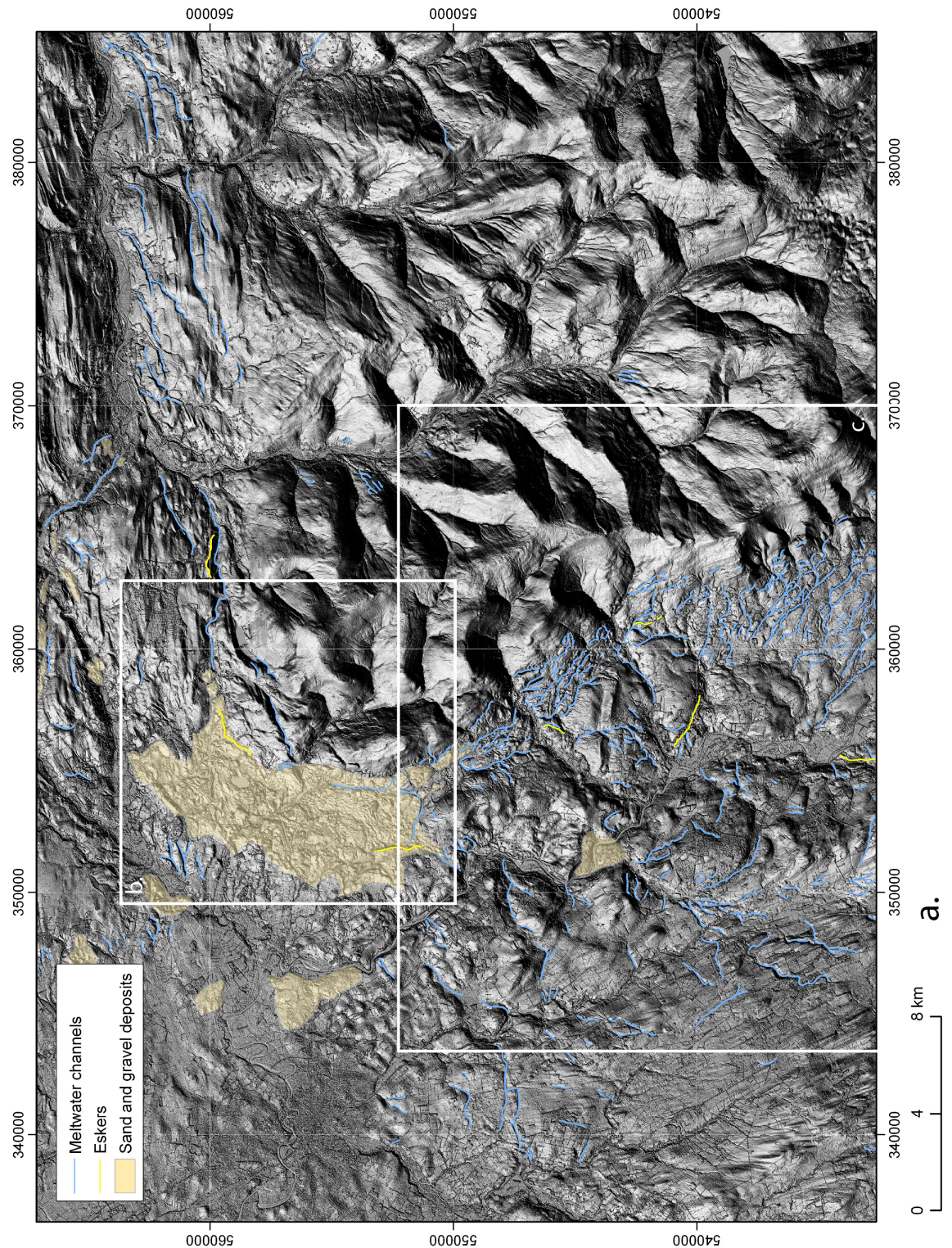
## 5.4 **Geomorphology**

The geomorphology of the area is displayed in Fig. 5.5a and can be sub-divided into four principal assemblages, namely ridges, depressions, flat-topped hills and meltwater channels.

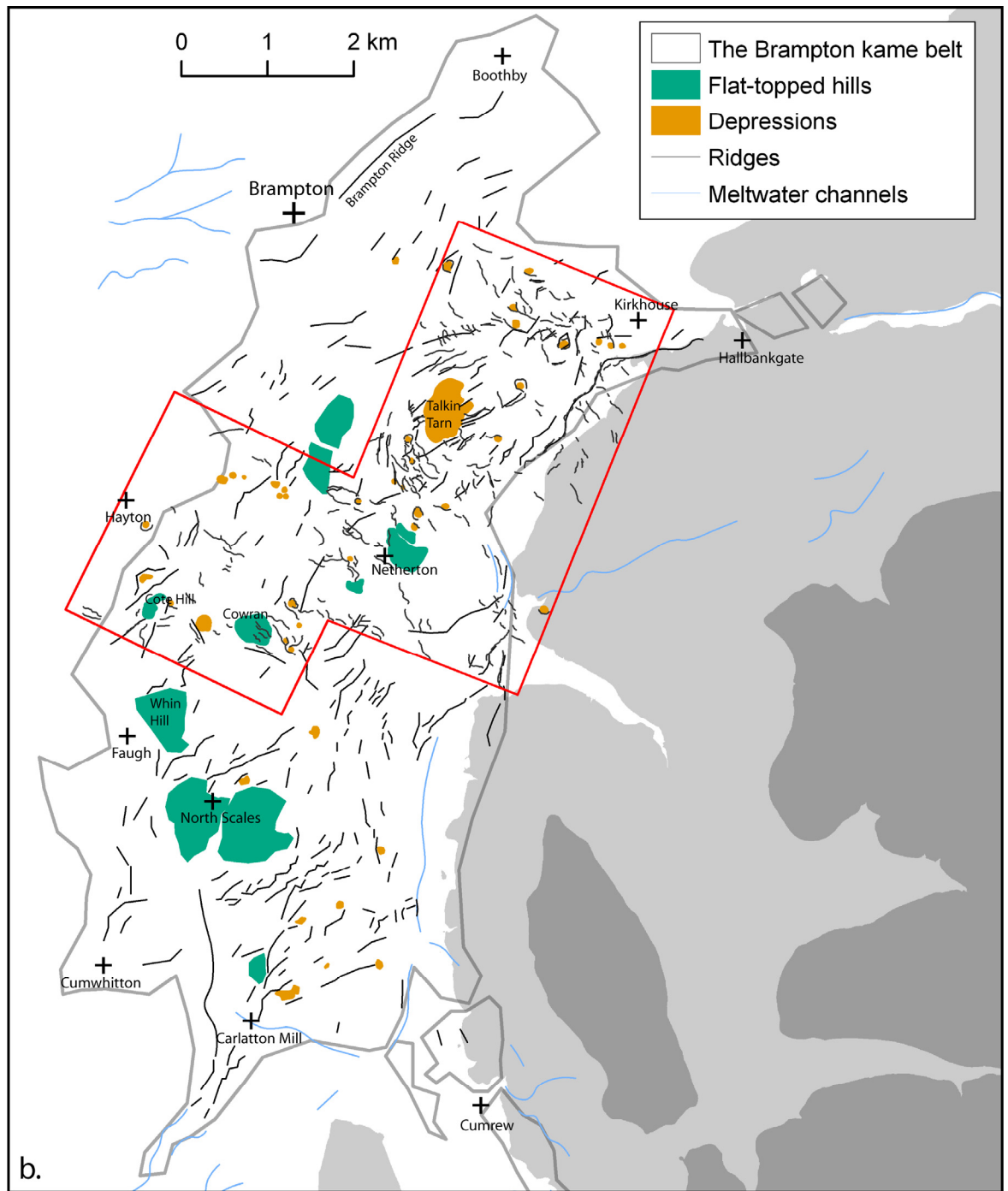
The first morphologically distinct feature comprises a complex system of ridges that occur throughout the entire kame belt. Most ridges wrap around the Pennine escarpment, trending SW-NE and then W-E around the northern-most section of the kame belt leading into the Tyne Gap (Fig. 5.5b). A secondary SE-NW trend of a more subdued series of ridges is apparent, especially in the aerial photographs. The ridges are discontinuous, with lengths ranging from less than 100 m to over 2 km and heights of generally no more than 20 m (Fig. 5.6). A larger, more prominent SW-NE aligned ridge at Brampton, to the north-west of the complex, rises to 50 m, is 370 m wide and extends for 1.5 km. Ridges tend to be smaller, more discontinuous and with higher levels of complexity to the south-east of the kame belt. In places the ridges are anastomosing, as in the region of Carlatton Mill but most are branched, becoming increasingly dendritic northwards.



Despite generally running along the contours of the slope, the ridges also display undulatory long profiles where they cross undulatory terrain. The ridges themselves are situated at various heights in the kame belt, ranging from less than 100 m O.D. at the westernmost end to 200 m O.D. where the Hallbankgate esker (Figs. 5.3 and 5.5b) winds into the Tyne Gap.







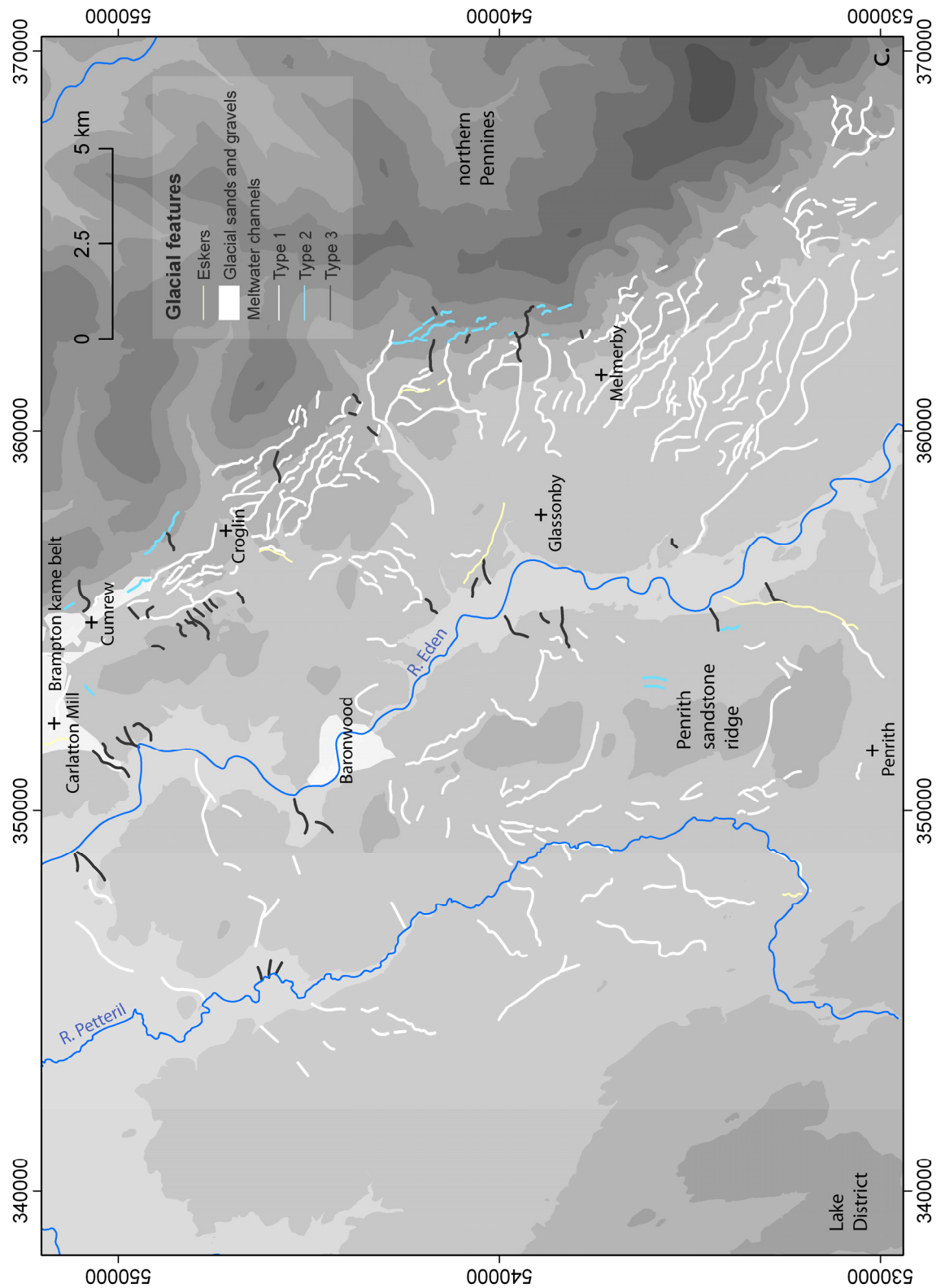


Figure 5.5: (a) NEXTMap image of the Brampton kame belt and surrounding area, with meltwater channels, eskers and glacial sand and gravel deposits identified; (b) glacial geomorphological map of the Brampton kame belt, flat-topped hills, ridges, meltwater channels and depressions; and (c) glacial geomorphological map of the meltwater channels, eskers and glacial sand and gravel deposits along the Pennine escarpment.

Second, a series of circular depressions can be found throughout the kame belt but are especially prominent in the south-east where the terrain appears pockmarked (Fig. 5.5b). They often punctuate ridges, thereby producing an undulating crest-line or discontinuous ridge form (Figs. 5.5b & 5.6). Depressions also occur on flat-topped hills. The depressions range from 20 m to 750 m in diameter, the largest being Talkin Tarn (Fig. 5.5b), although most are between 20 m and 60 m wide.

Flat-topped hills up to 1 km wide and 20 m high make up the third landform assemblage. The biggest of these occur at Stonebridge, Whin Hill, Cowran, Cote Hill, Netherton and just west of Talkin Tarn (Figs. 5.1 & 5.5b). Flat-topped hills are generally associated with ridge networks which are sometimes partially buried by the hills. Circular depressions are also common on the flat-topped hills. The flat-topped hills tend to occur within a central zone of the Brampton kame belt measuring 1.5 km wide by 6 km long, where they lie in close proximity to each other (Fig. 5.5b) and range in height from 105 m O.D. to 140 m O.D.



Figure 5.6: Oblique aerial photograph (Cambridge University) looking north-west along the edge of Talkin Tarn, revealing a series of subdued ridges (marked x).

Finally, meltwater channels occur but are generally rare within the kame sequence, although a few south-north orientated channels exist on the lower slopes of the Pennine escarpment. Another

channel cuts SE-NW towards the bottom of the kame belt at Carlatton Mill (Fig. 5.5b, c). A total of 253 channels were mapped along the length of the Pennine escarpment. On the south-eastern margin of the Brampton kame belt a series of SE-NW orientated channels cut into the kame topography giving it a ridged appearance (Fig. 5.5c). This forms the northern edge of a major meltwater system running along the edge of the Pennine Escarpment and stretching as far south as the Stainmore Gap (Figs. 5.5c) (cf. Livingstone *et al.*, 2008). There are three major channel forms within the drainage network (Figs. 5.5c & 5.7) all of which cut into the Permian and Triassic bedrock. Type 1 channels are the most common channel form (a total of 205 channels mapped), are typically 'U' shaped, deeply incised and continuous, and can be traced for over 4 km in some instances, with a mean length of 995 m (Figs. 5.5c & 5.7a). They also exhibit undulatory long profiles. They form an anastomosing network of roughly parallel channels trending SE-NW towards and into the Eden Valley and are situated between 100-450 m O.D. (Livingstone *et al.*, 2008). Type 2 channels are rarer (23 mapped channels), generally smaller (< 1 km, with a mean of 457 m), trend perpendicular to the hillslope gradient and are orientated in a more northerly direction. These channels lack the undulations seen within the Type 1 channels and are found higher up the slopes of the Pennine escarpment (generally between 300 and 450 m O.D.) (Fig. 5.5c). Finally, Type 3 channels are small (< 400 m) and deeply incised and plunge westwards, from inception points generally associated with the Type 2 channels (Figs. 5.5c & 5.7b). A total of 27 channels fall within this category and they all have steep gradients.

A further 100 meltwater channels have been identified criss-crossing the Penrith sandstone ridge (up to 240 m O.D.), a prominent outcrop in the Vale of Eden (Fig. 5.5c). Three main channel forms consistent with those of the Pennine escarpment are again recognised. A total of 81 type 1 meltwater channels are identified on the ridge (Fig. 5.5c). This includes a dendritic pattern of channels that trend north-eastwards into a wedge of glacial sands and gravels at Baronwood (interpreted as an ice-contact delta by Huddart, 1970) (Fig. 5.5c). Type 1 meltwater channels also form dendritic networks with eskers on the southern and eastern edge of the ridge (Fig. 5.5c). Type 2 channels are less common (16 mapped) and are characterised by short lengths either plunging down gradient into the Eden Valley off the Penrith sandstone ridge or into the Eden Valley from the Brampton kame belt. Type 3 channels are rare (3 mapped) within this region and incised at a much lower elevation (<150 m O.D.) than the Type 3 meltwater channels of the Pennine escarpment (Fig. 5.5c). The channels mapped on the Penrith sandstone ridge are in general isotropic, although the western side of the ridge is dominated by SW-NE orientated channels (Fig. 5.5c).

The north-eastern arm of the Brampton kame belt feeds into (via the Hallbankgate esker) an 8 km long, deeply incised meltwater channel that cuts across the watershed into the Tyne Valley system (Fig. 5.5a). The channel possesses undulations in the western end as it crosses the watershed, before gently dipping down into the South Tyne Valley.





a.



b.

Figure 5.7: (a) Photograph showing a typical Type 1 meltwater channel from the Pennine escarpment; and (b) photograph showing a typical Type 3 meltwater channel which plunges down the Pennine escarpment and is deeply incised.

## 5.5 Sedimentology and stratigraphy

### 5.5.1 Ridges

Several exposures have been logged through ridges within the Brampton kame belt. These include a quarry at Kirkhouse near Milton, two sites used by Huddart (1970) at Brampton ridge and Quarrybeck and a site north of Whin Hill. These exposures are augmented with borehole data collected by the British Geological Survey (Jackson, 1979).

#### 5.5.1.1 Kirkhouse

Kirkhouse sand and gravel quarry (NY 563 601) is located east of Milton in the northern kame belt region (Fig. 5.1). The western end of the quarry leads into a NW-NE orientated ridge, 300 m long and 140 m O. D. The quarry contains 4 major lithofacies associations (LFA) which are outlined diagrammatically in Figure 5.8 and are described and interpreted in turn below.

#### Lithofacies Association 1 (LFA K1; series of gravel channel fills and trough cross-bedded sand and gravel) - description:

Lithofacies association 1 (LFA K1) is characterised by 0.2 - 1.5 m thick tabular cross-beds and planar beds of fine-coarse sand, a series of vertically stacked trough cross-beds ranging from 0.5 - 5 m in width and clast-supported gravel channels up to 0.5 m thick and 5 - 10 m wide (Figs. 5.8, 5.9a,b). The trough cross-beds exhibit fining upwards sequences from stratified coarse sand and granule gravel (with some pebble gravel) into fine-medium sands. These structures often cut-across the tabular cross-beds and planar beds of fine-coarse sand (Fig. 5.9a,b) which contain outsized clasts. The gravel channels, which are generally found towards the bottom of the LFA, have an erosional base and are predominantly structureless apart from infrequently observed centimeter-scale lenticles of sand. LFA K1 is rarely found throughout the quarry, with sequences generally vertically stacked (up to 10 m thick) between and within LFA K2 and K3. Bounding surfaces are characterised by marked shifts in grain size, bedform scale and lithofacies type, with frequent erosional surfaces.

#### LFA 1 - interpretation

The stacked multi-storey sequences of LFA K1 indicate that vertical accretion was the prevalent depositional mechanism (Collinson, 1996). The sets of trough cross-beds result from the filling of channels by migratory sand dunes in a low energy environment (Miall, 1977), while planar-bedded sand demonstrates transport as a traction carpet in the upper-flow regime (Miall, 1977; Allen, 1984). Gravel at the base of the troughs is associated with erosional re-activation of the facies, with upwards fining eventually leading to abandonment (Smith, 1985). Pebble-gravel channels indicate bedload transport, possibly as channel lags or channel bars, from high energy stream flows (Collinson, 1996). The lithofacies association is therefore indicative of a fluvial environment



characterised by aggradation and then abandonment of channels and reflecting channel, dune and bar migration in a fluctuating hydrological regime typical of braided rivers (e.g. Williams & Rust, 1969; Rust, 1972; Miall, 1977).

#### Lithofacies association 2 (LFA K2; laminated sands and ripple structures) - description

LFA K2 (S2A & B, Fig. 5.8) comprises a series of fine sand laminations characterised by fining-upwards sequences, ripple structures and out-sized granule-pebble gravel (Fig. 5.9c). This lithofacies association is infrequently observed and generally interbedded within LFA K1 and K3. Thicknesses rarely exceed 2 m whilst bed widths are generally greater than 10 m. Ripples are typically found at the base (and infrequently at the top) of the lithofacies and include: gently sloping ( $< 10^\circ$ ), laterally discontinuous type 'A' climbing ripples (Jopling & Walker, 1968); and fine-grained sinusoidal ripples up to 5 cm thick (Fig. 5.9c). Palaeo-current directions are variable, although the general trend is towards the NNE (S2B, Fig. 5.9c).

#### LFA 2 - interpretation

The horizontally laminated sand is interpreted to have formed by low energy traction currents during planar bed flow (Jopling & Walker, 1968; Smith & Ashley, 1985). The development of type 'A' climbing ripples, at the top and base of the lithofacies association, demonstrates the waxing and waning flow of water respectively (Miall, 1985), with an increase in suspended sediment input and a reduction in flow velocity relative to the horizontal laminations. Fine-grained sinusoidal ripples demonstrate deposition predominantly by suspended sediment rain-out in very quiet waters (Jopling & Walker, 1968; Allen, 1973; Ashley *et al.*, 1982). The palaeocurrent data suggests that flow generally trended NNE, although with significant variability. Thus the LFA is indicative of a low energy environment characterised by variations in the suspended sediment input and flow velocity.

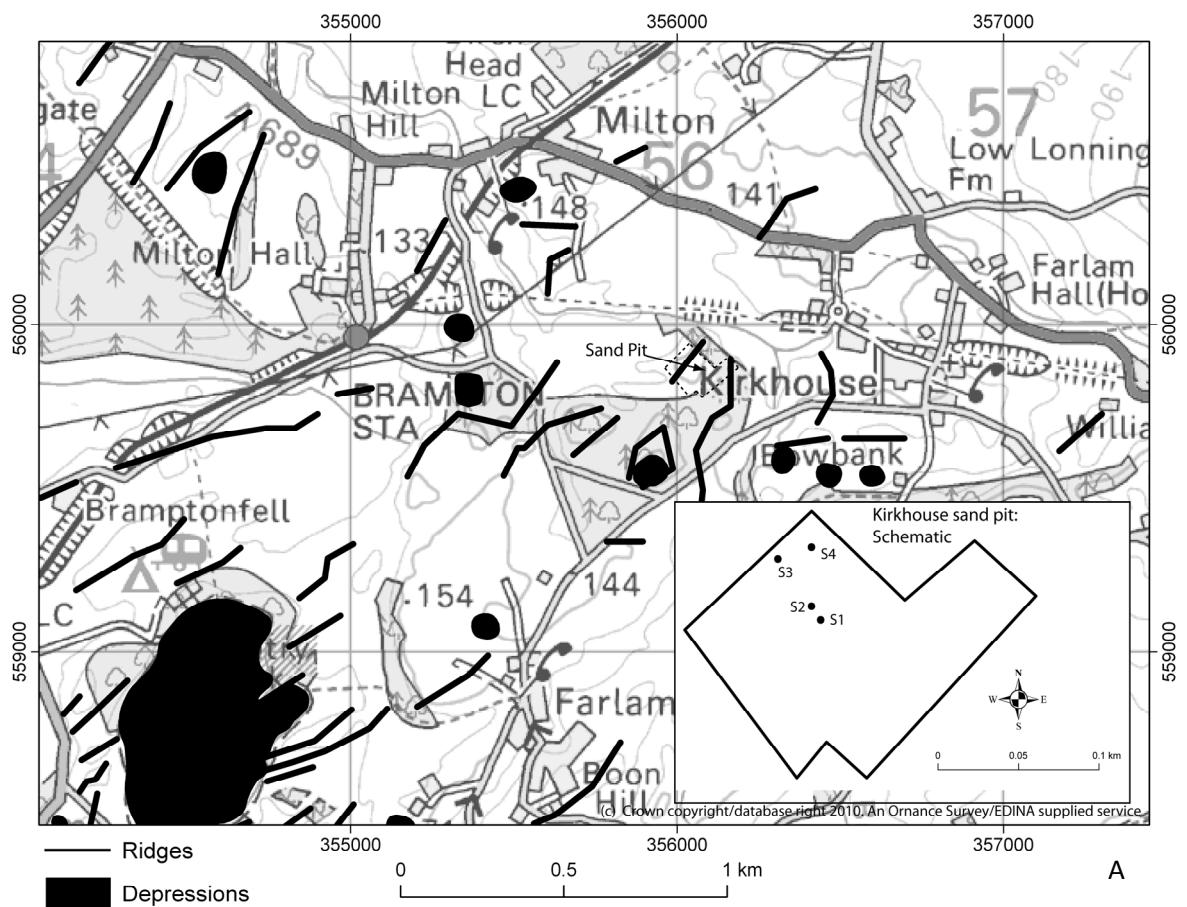
#### Lithofacies Association 3 (LFA K3; series of tabular cross-bedded stratified sand and fine gravel, draped clay/silt bands and pebble gravel) - description

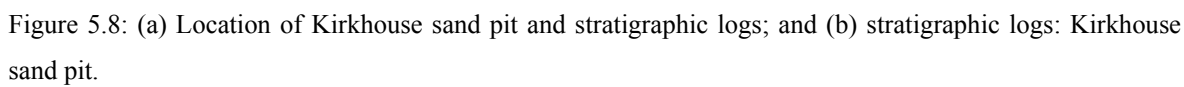
LFA K3 (S1, 2B, 3 & 4, Fig. 5.8) is composed of tabular cross-beds of stratified medium-coarse sand and granule gravel with drapes of clay/silt and occasional pebble gravel lithofacies (Fig. 5.9d). Stratified medium-coarse sand and granule gravel lithofacies are *ca.* 0.8 m thick and dip at various orientations from the north to the northeast at angles ranging from 18-30°. Each dipping unit is bounded by an erosional re-activation surface. Massive, coarse sand containing granule-pebble gravel up to 0.5 m thick is often observed (Fig. 5.9d), whilst there are also rare, shallow ( $< 1$  m) granule-pebble gravel lithofacies that display trough and lenticular geometries and erosional lower

surfaces. Occasional fine-grained drapes up to 15 cm thick and with undulatory profiles are observed at irregular intervals throughout the lithofacies association (Fig. 5.9d), with the greatest frequencies (and thicknesses) at the eastern end of the quarry. LFA K3 is the dominant sedimentary association within the quarry reaching thicknesses of up to 15 m and extending throughout the quarry.

### LFA 3 - interpretation

The tabular cross-bedded stratified sand and granule gravel lithofacies are interpreted as dune forms (Miall, 1977) analogous to the deeper portions of active channels where the bedload is predominantly sand; the lateral extent and range in orientations indicative of a broad, unconfined channel (Miall, 1985). Vertical accretion of discrete co-sets bounded by re-activation surfaces is typical of migratory dune formation (Banerjee & McDonald, 1975). The interbedded coarse sand with granule-pebble gravel could be interpreted as antidunes (Shaw, 1972), thus suggesting that deposition occurred sub-aerially (Banerjee & McDonald, 1975). The irregularly spaced clay bands were formed by the settling out of fines, which therefore demonstrates evidence for the periodic, pervasive shut-down of the fluvial system (Miall, 1977). These dune bedforms, which show evidence for continual re-activation, migration and vertical accretion, could form in a low flow, braided river system characterised by fluctuating discharge (Miall, 1977).





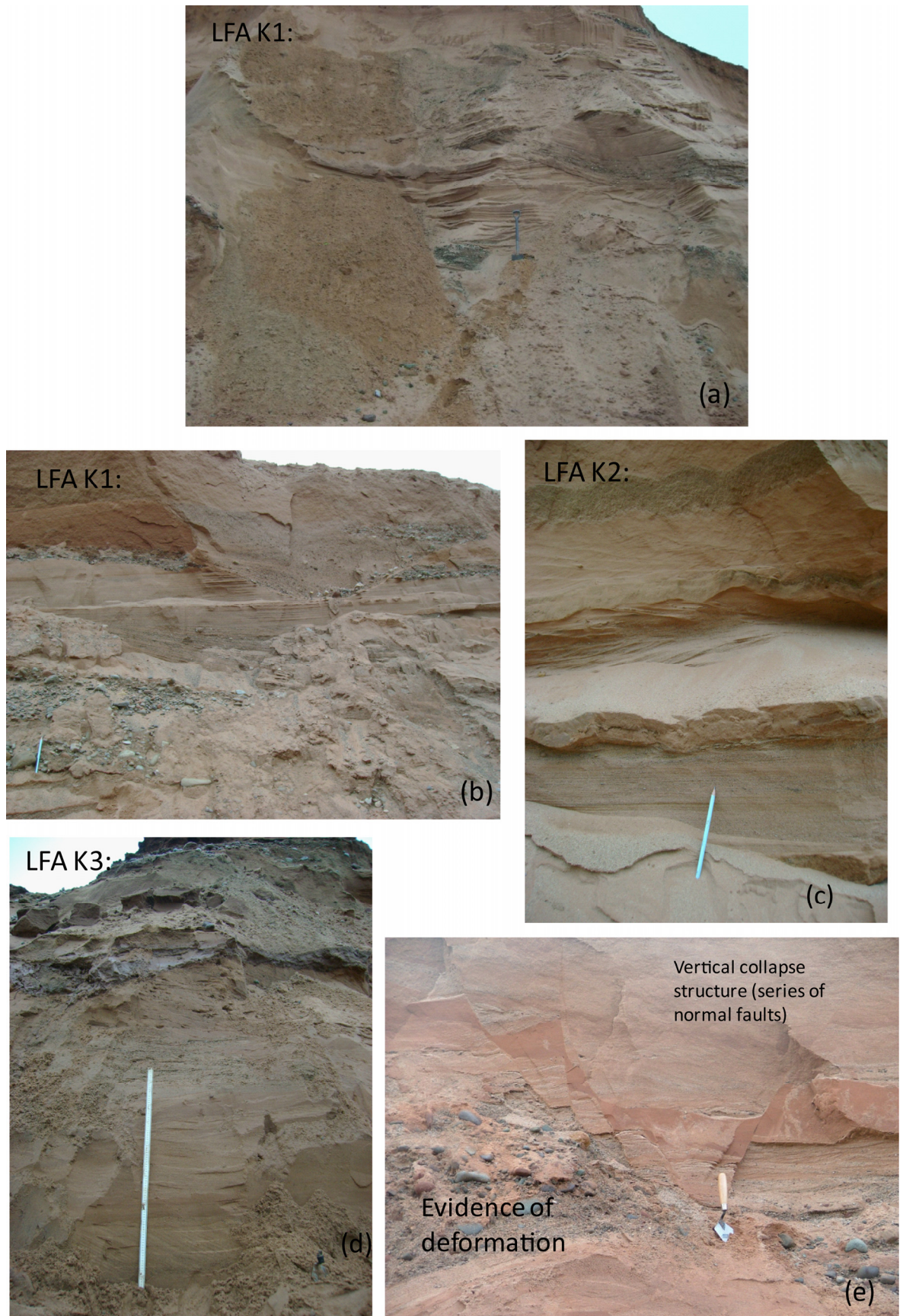


Figure 5.9: Photographs from Kirkhouse sand pit: (a) LFA K1: tabular cross-beds and planar beds of fine-coarse sand and a couple of trough cross-beds comprised of stratified coarse sand and granule gravel; (b) LFA K1: scour trough cut into parallel laminated sand and comprised of stratified sand and granule gravel <

Fig. 5.9 (*continued*)... with some pebble gravel at the base; (c) LFA K2: thin alternating facies of planar laminated fine sand, fine-grained sinusoidal drapes and type 'A' climbing ripples; (d) LFA K3: tabular cross-bedded sand, stratified coarse sand and granule gravel, massive coarse sand and fine-grained drapes; and (e) vertical collapse structure.

#### Evidence for deformation - description

Deformation is persistent throughout LFA K1, 2 and 3 with the most extensive evidence at the base of the vertical sequence (S1, 2 & 4, Fig. 5.8). This includes 5 - 10 m wide by 2 - 5 m deep open folds (S1, 2 & 4, Fig. 5.8) characterised by high angle reverse faults (with offsets of up to 5 cm), overlapping facies, over-steepened stratified sand and granule gravel, truncated facies, and a series of normal faults (millimeter - centimeter scale) along the crest of the folds (S2, Fig. 5.8), often arranged *en echelon*. Indeed normal faulting is common throughout the quarry, often characterised by block failure (Fig. 5.9e), while there is also extensive evidence of centimeter-scale incorporation, soft sediment pods, stringers, ball and pillow, water escape and flame structures (S1, Fig. 5.8), and displaced blocks of sand within the stratified sand and granule gravel.

#### Evidence for deformation - interpretation

Vertical deformation structures such as the open folds and block failures are associated with collapse of the depositional floor owing to the melt-out of buried ice blocks (McDonald & Shilts, 1985; Krüger, 1994; McCarroll & Rijdsdijk, 2003) at a variety of scales. The preponderance of normal faults and displaced blocks of sediment is indicative of gravitational failure with the displaced blocks analogous to bank collapse. High pore waters must have existed either during or following deposition due to the presence of stringers, water escape, ball and pillow and flame structures.

#### Kirkhouse Ridge synopsis: environment of deposition

The sedimentary lithofacies associations at Kirkhouse sand and gravel quarry (Figs. 5.8, 5.9) are envisaged to have been deposited in a broad, unconfined, glaciofluvial braided channel resting on dead ice (cf. Price, 1966; Bennett & Glasser, 1996; Huddart *et al.*, 1999). The braidplain consisted of co-sets of dunes (sand flats) associated with unsteady, unconstrained flow undergoing periodic re-organisation and occasional abandonment (LFA K3), and regions of more persistent flow (LFA K1) where migratory channels dominated (cf. Nemec, 1992). The complex shifts in grain size, facies and lithofacies associations and the reactivation surfaces are indicative of glaciofluvial outwash undergoing fluctuating discharge, channel migration and pulsed incision (e.g. Owen, 1997). The prevalence of dead-ice structures and outsized clasts (which in this context were



probably deposited as drop-stones) throughout all the lithofacies associations suggests that the glaciofluvial outwash was in contact with ice (Shaw, 1972; Huddart *et al.*, 1999). Bank failure blocks and gravitational slumping indicate that the environment was very unstable, whilst the presence of ball and pillow and flame structures demonstrate mass-flowage and collapse of the upper facies due to dead ice disintegration, leading to the loading and de-watering of fine grained material (cf. Cheel & Rust, 1982; Donnelly & Harris, 1989; Mager & Fitzsimons, 2007).

#### 5.5.1.2 Brampton Ridge (BR), Quarrybeck Pit (Huddart, 1970)

### Description

Brampton ridge forms the most prominent feature of the kame belt, running north-east from Brampton for 0.9 km and rising 50 m, with peaks of 120 m O. D. on its undulating crest (Figs. 5.1, 5.5). The ridge is located in the northerly-most region of the kame belt with the quarry situated in the very north-eastern edge of the ridge (NY 543 623). Huddart (1970) collected 3 logs in the exposed face, which are reproduced here in Figure 5.10.

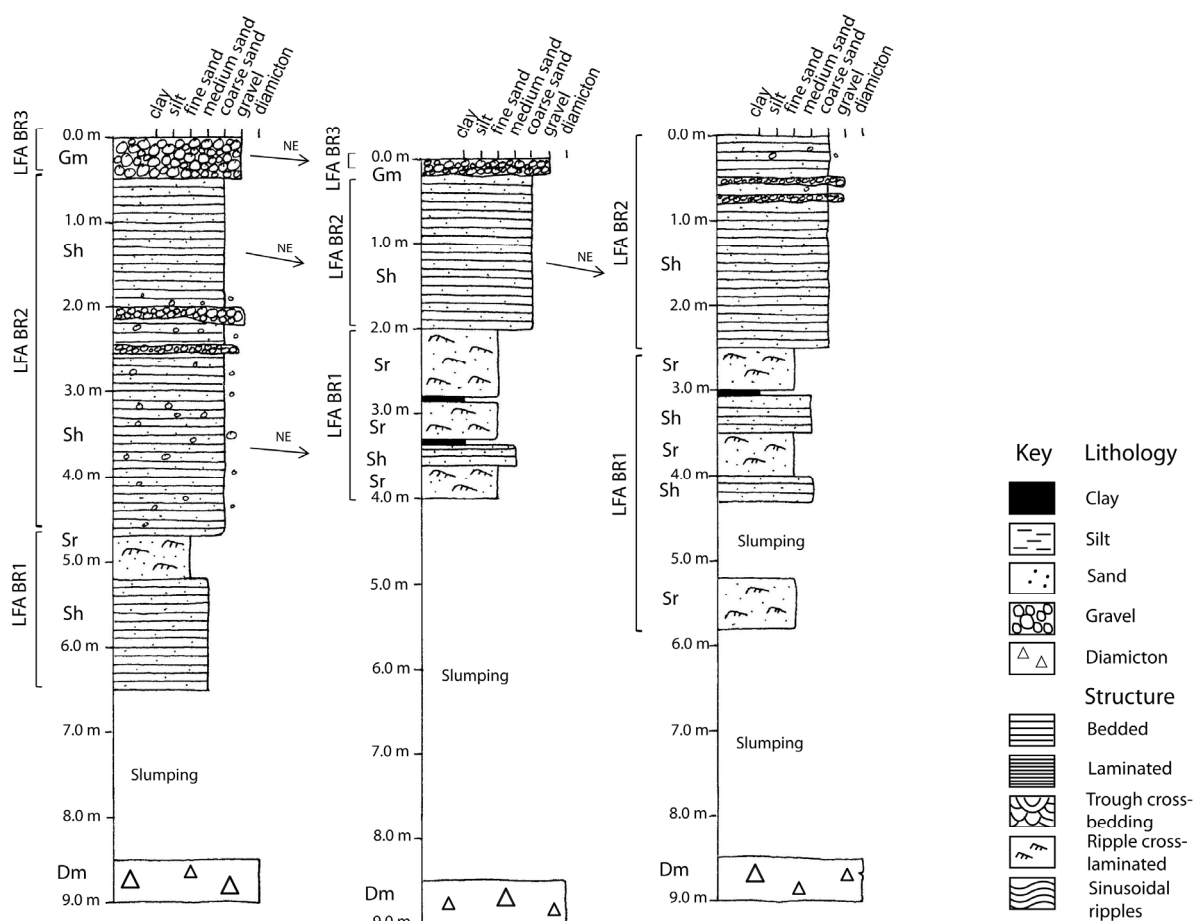


Figure 5.10: Stratigraphic logs: Brampton ridge sand pit (reproduced from Huddart, 1970).



The stratigraphy is composed predominantly of sand, but with a general upwards coarsening into a pebble-gravel facies at the top. Below the currently exposed sequence Huddart (1970) observed red clay which he interpreted as basal till (Fig. 5.10). Three major lithofacies associations were observed within the sequence. LFA BR1 comprises rippled fine sand with a maximum thickness of 35 cm, interbedded with horizontally stratified medium-coarse sand dipping at 5° to the north-east, thin clay bands and silt layers. LFA BR2 comprises 2.66 m of horizontally stratified, coarse sand with thin pebble layers and erosional scours. Cross-stratification is also evident in the lower part of LFA BR2, with foreset dips of 10° and a palaeo-current indicating deposition from the south-west (Huddart, 1970). The final lithofacies (LFA BR3) is composed of 55 cm of pebble-gravel dipping at 5°.

#### Interpretation

Deposition of horizontally stratified sand and ripple structures at the base of the sequence is indicative of a low energy flow regime (Collinson, 1996). The horizontally stratified sand is envisaged to have formed by planar-bed flow in shallow water (hence preventing dune development), possibly as a result of sheet flooding (Miall, 1985; Collinson, 1996). The progressive coarsening up-sequence into pebble gravel is indicative of bar and channel development in the upper flow regime, and palaeocurrents suggest that flow was parallel with the north-east orientation of the ridge. The exposure at Brampton ridge is therefore interpreted to have been deposited in a fluvial channel, which, because it is located in an upstanding ridge, must have been ice-walled (cf. Price, 1969, 1971, 1973).

##### *5.5.1.3 Braithwaite's sand and gravel pit (Huddart, 1970)*

#### Description

Braithwaite's sand and gravel pit (NY 512 568) is situated in a 300 m SW-NE trending ridge north of Whin Hill (Figs. 5.1 and 5.5). Huddart logged both the lateral and vertical extent of the ridge (Fig. 5.11). The central axis of the ridge (logs 5 and 6) grades up through a red, sandy clay lithofacies containing diamicton balls into 4 m of large scale, trough cross-stratified coarse sands, 5 m of large pebble-gravel filled channels, 2.3 m of imbricate pebble-gravel and 3 m of horizontally stratified coarse sands. Exposures on either side of the ridge axis are dominated by alternations of horizontally stratified coarse sand, tabular cross-stratified sand, fine grained bands and sinusoidal and type 'A' ripple-drift cross-laminated sand (Jopling & Walker, 1968). The tabular cross-stratified lithofacies have foreset dips ranging from 22 - 26°. Marginal faults dipping towards the south-southeast are evident below the top unit of pebble-gravel, between logs 5 and 7 in the marginal zone of the ridge. Palaeocurrents are varied, ranging from east to north-east in the faulted gravel, to west in the tabular cross-stratified sand.

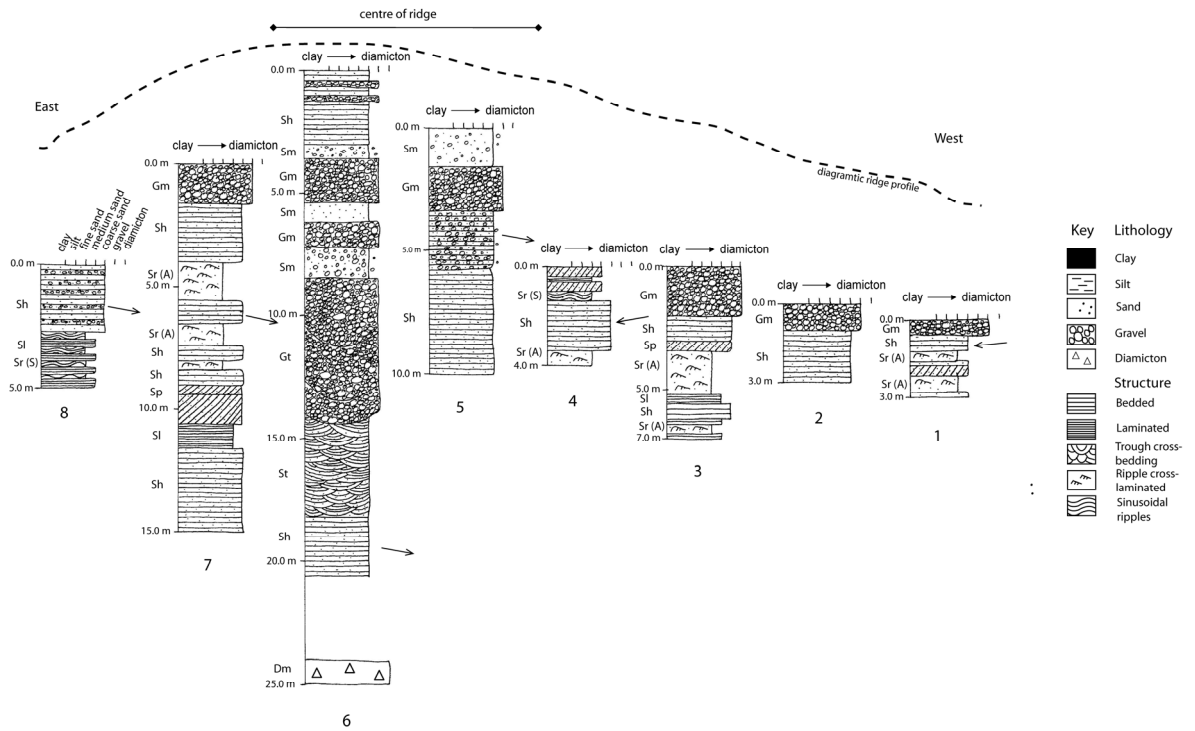


Figure 5.11: Stratigraphic logs: Braithwaite's sand pit (reproduced from Huddart, 1970).

#### Interpretation:

The exposures through the central axis of the SW-NE trending ridge are indicative of a high energy flow regime, where trough cross-stratified coarse sands represent dune forms (Miall, 1977). The horizontal sand is interpreted as an upper flow regime shallow sheet flood deposit. The gravel lithofacies represent either channel scour or bar deposits, with imbricate gravels indicating traction deposition (Miall, 1977, Collinson, 1996). Thus the central exposures record fluvial deposition typical of a braided stream (e.g. Miall, 1977) characterised by fluctuating, unsteady and high energy flow (Collinson, 1996). The deposits towards the ridge margin indicate deposition in the lower flow regime, with shallow angled tabular cross-stratified sand thought to represent dune formation in the lower flow regime (Miall, 1977). The horizontal coarse sand is interpreted as a sheet flood deposit (Collinson, 1996) and alternations of sinusoidal ripples and fine-grained drapes are analogous to suspended sediment rain out. Type 'A' climbing ripples record a relative reduction in suspended sediment and an increase in bedload transport (Jopling & Walker, 1968; Allen, 1973; Ashley *et al.*, 1982).

##### 5.5.1.4 Boreholes: Mineral Assessment Report 45

Borehole data (Jackson, 1979) through ridges at Hall Bank (NY 551 588) and Tarn End (NY 543 580) (Fig. 5.1) provide additional stratigraphic and sedimentological information from the

depositional ridges (Fig. 5.12). A red-grey till containing shale and sandstone is evident in the bottom 2.1 m of Hall Bank borehole. Sands and gravels within the ridges are up to 13.2 m thick. At Hall Bank the basal 1.2 m comprise gravel and occasional boulders, followed by 1.4 m of massive red sandy silt and then 6 m of sand (Fig. 5.12). This documents an abrupt change in discharge regime during the early stages of ridge deposition. Deposits at Tarn End comprising more than 6.5 m of pebbly sand followed stratigraphically by 6.7 m of gravel (with clay at the very top) are indicative of a high energy fluvial system.

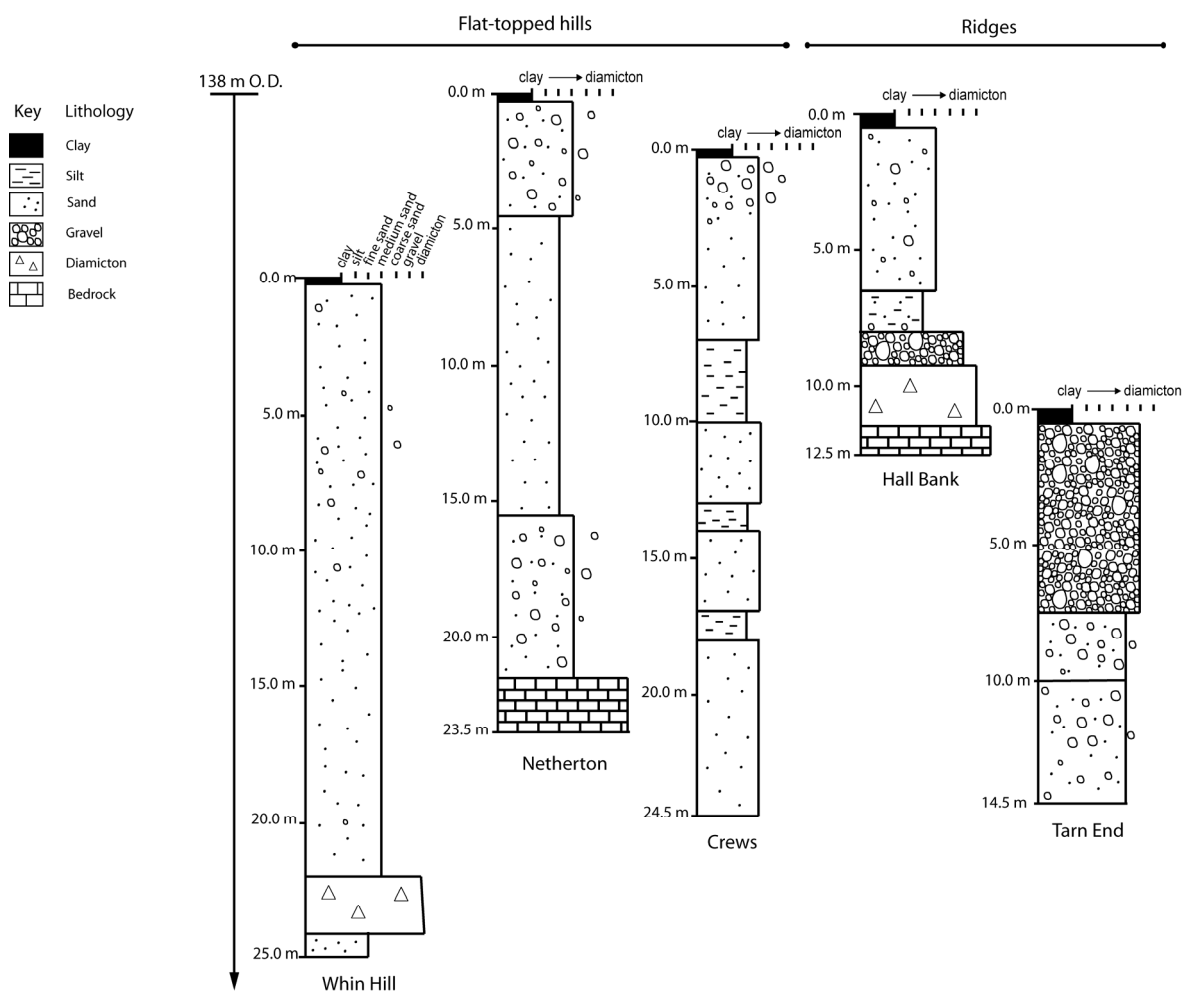


Figure 5.12: Borehole logs through several ridges and flat-topped hills, reproduced from Jackson (1979).

#### Synopsis of quarry exposures, and borehole data in ridges: environment of deposition

Stratigraphic and sedimentological evidence collected through ridges in the Brampton kame belt can be reconciled with a fluvial regime characterised by unsteady, migratory flow (cf. Miall, 1977, 1985; Owen, 1997). Marginal faults at Braithwaite's sand and gravel pit suggest that the ridge was in contact with an ice mass (Huddart, 1970; Shaw, 1972; Gustavson & Boothroyd, 1987; Warren & Ashley, 1994). Marginal faults are envisaged to have been formed by gravitational collapse due to the removal of a supporting ice wall (McDonald & Shilts, 1985). A typical trend seems to be that

of a central channel forming along the axis of the ridge characterised by vertical accretion of dune forms and incised channels in a high energy, unsteady flow regime, together with shallow sheet flow, characterised by periodic channel abandonment, and water depth fluctuations along the ridge margins. The presence of both normal and reverse grading is a product of a highly variable discharges, whereby continual abandonment and reactivation results in some channels becoming major conduits (coarsening upwards), whilst others devolve into small backwater outlets (fining upwards). Similar lithofacies associations in sinuous ridges have been previously reported from eskers (e.g. Shaw, 1972; Warren & Ashley, 1994), wherein esker cores often display cyclic sequences of gravel and sand (Bannerjee & McDonald 1975; Ringrose 1982; Brennand 1994; Brennand & Shaw 1996). Fining-upward sequences with erosive bases record fluctuations in discharge and sediment availability, and the depositional cycles (rhythmicity) have been interpreted as both annual (Bannerjee & McDonald 1975; Mäkinen 2003) and seasonal (Brennand 1994).

Horizontal facies variability is also common in situations where eskers contain cavity fill sequences, produced when tunnels widened, either due to flotation in proglacial lakes (e.g. Gorrell & Shaw 1991; Brennand 1994, 2000) or leakage of tunnel waters into lateral, ephemerally water-filled cavities (e.g. Gordon et al. 1998; Mair et al. 2002). The latter have been linked to the construction of anabranching reaches, subaqueous fans and hummocky zones along otherwise predominantly single ridged eskers by Brennand (1994, 2000). Fining outwards from the main ridge is thought to represent side-wall melting and esker widening, an interpretation supported by the occurrence of faulting. An esker may be “widened” by sediment draping where discharge from the tunnel portal is into a subaqueous environment, which may be at the flotation point in a proglacial lake or where the tunnel exits into a supraglacial pond.

### 5.5.2 *Flat-topped hills*

Several exposures have been logged through flat-topped hills (Figs. 5.1 and 5.5) within the Brampton kame belt. This includes a sand pit at Faugh, a site at Whin Hill (Huddart, 1970) and borehole data collected by the British Geological Survey (Jackson, 1979).

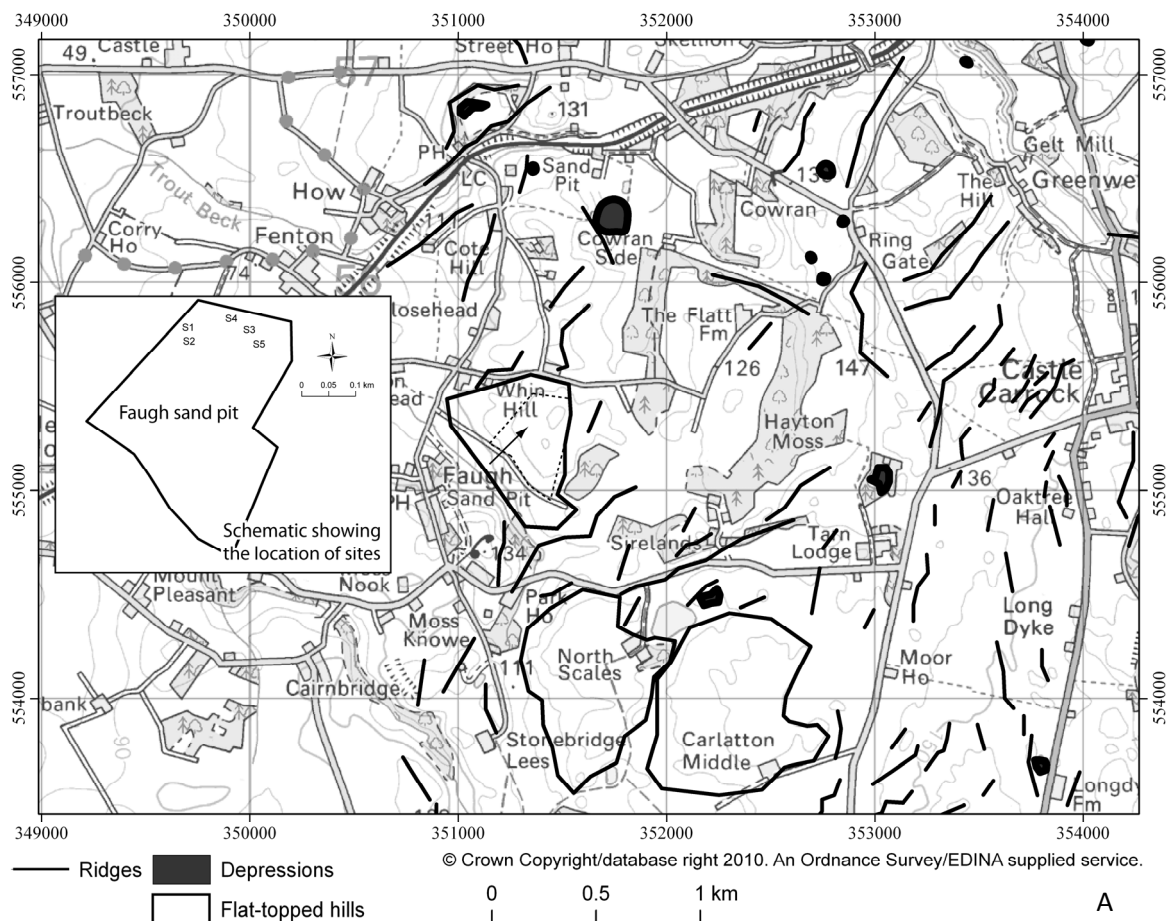
#### 5.5.2.1 *Faugh sand pit:*

Faugh sand pit (NY 514 550) is located at Whin Hill (120 m O.D.), to the northeast of Faugh village. Whin Hill is a 710 m long and 590 m wide flat-topped hill standing up to 20 m above the surrounding landscape. Ridges, orientated northwards, run into and out of the hill (Fig. 5.5). Four major lithofacies associations have been identified in the stratigraphic section and these are outlined diagrammatically in Figure 5.13.

Lithofacies Association 1 (LFA F1; series of fine sands, silts and clays) - description

At the base of the sequence in the north-western corner of the sand pit (S1, Fig. 5.13), and also higher up in the northern section of the pit (S4, Fig. 5.13), facies are dominated by alternations of horizontally bedded clay, silt and fine sand, massive fine-medium sand (Fig. 5.14a) and type 'B' climbing ripples, which often grade up into sinusoidal climbing ripples (Jopling & Walker, 1968). The climbing ripples dip at an angle of  $18^\circ$ , with a general orientation of  $322^\circ$  (palaeocurrent towards the north-west). Individual facies vary in thickness from as little as 5 cm to over 1 m, whilst the overall thickness of the sequence is *ca.* 6 m thick. Evidence of deformation includes centimetre scale folding, faulting and incorporated boundaries.

Further to the south (S2, Fig. 5.13), exposures reveal 4 m of WNW dipping ( $30^\circ$ ) sand, parallel laminated clay and sand, massive sand containing granule-gravel, and silt and clay bands up to 30 cm thick (Fig. 5.14b). In the bottom 50 cm of the exposure a series of clay, silt and sand laminations are truncated by stratified sand. The dipping sands are bounded by horizontal clay bands. There is evidence for deformation throughout the S2 exposure, with incorporated, convoluted silt and sand in the lower 50 cm, and a series of centimetre-scale fault structures and warped silt layers in the upper sequence. Just to the south-east of the exposure a meter-scale normal fault is also evident.



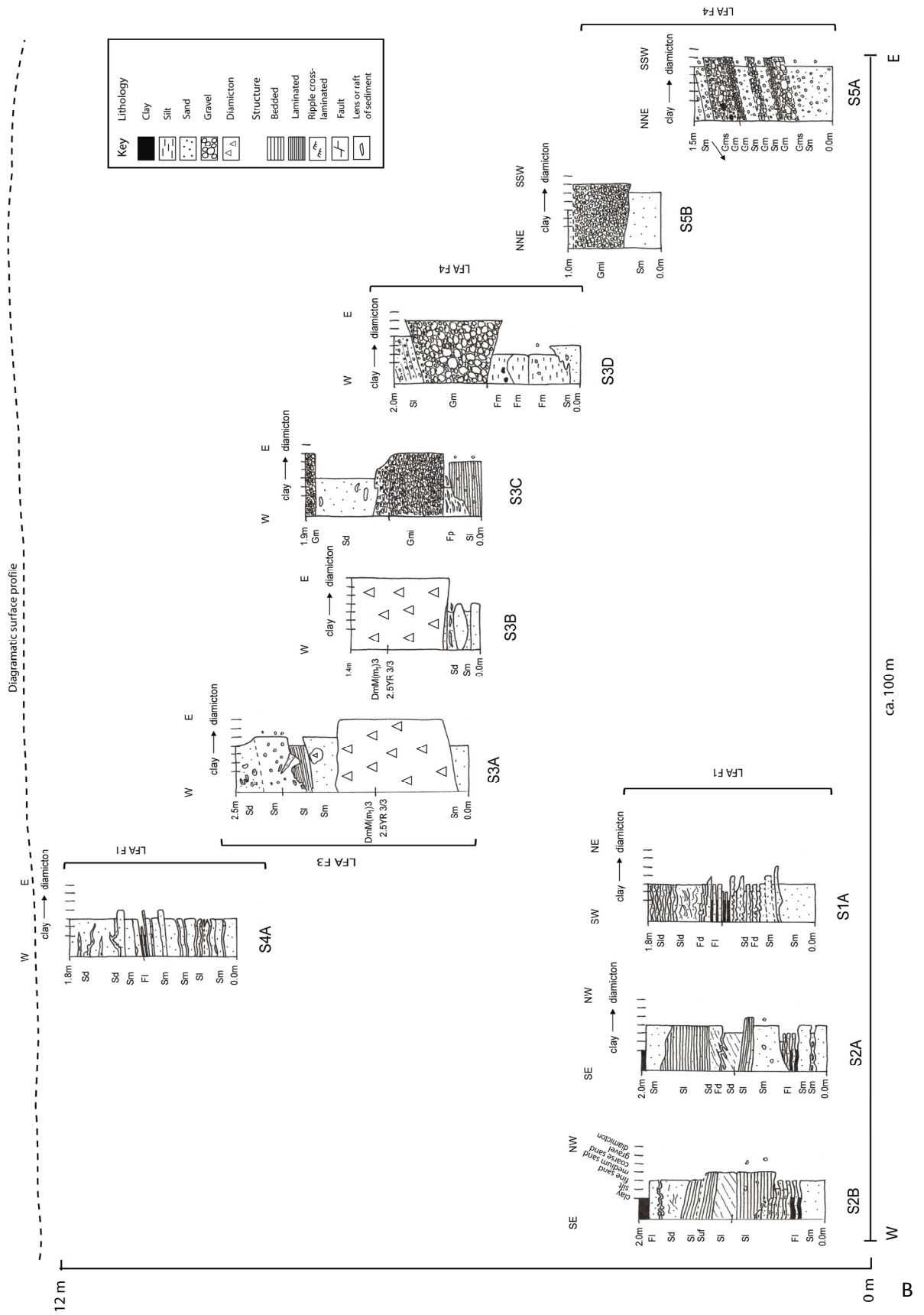


Figure 5.13: (a) Location of Faugh sand pit and stratigraphic logs; and (b) stratigraphic logs: Faugh sand pit.



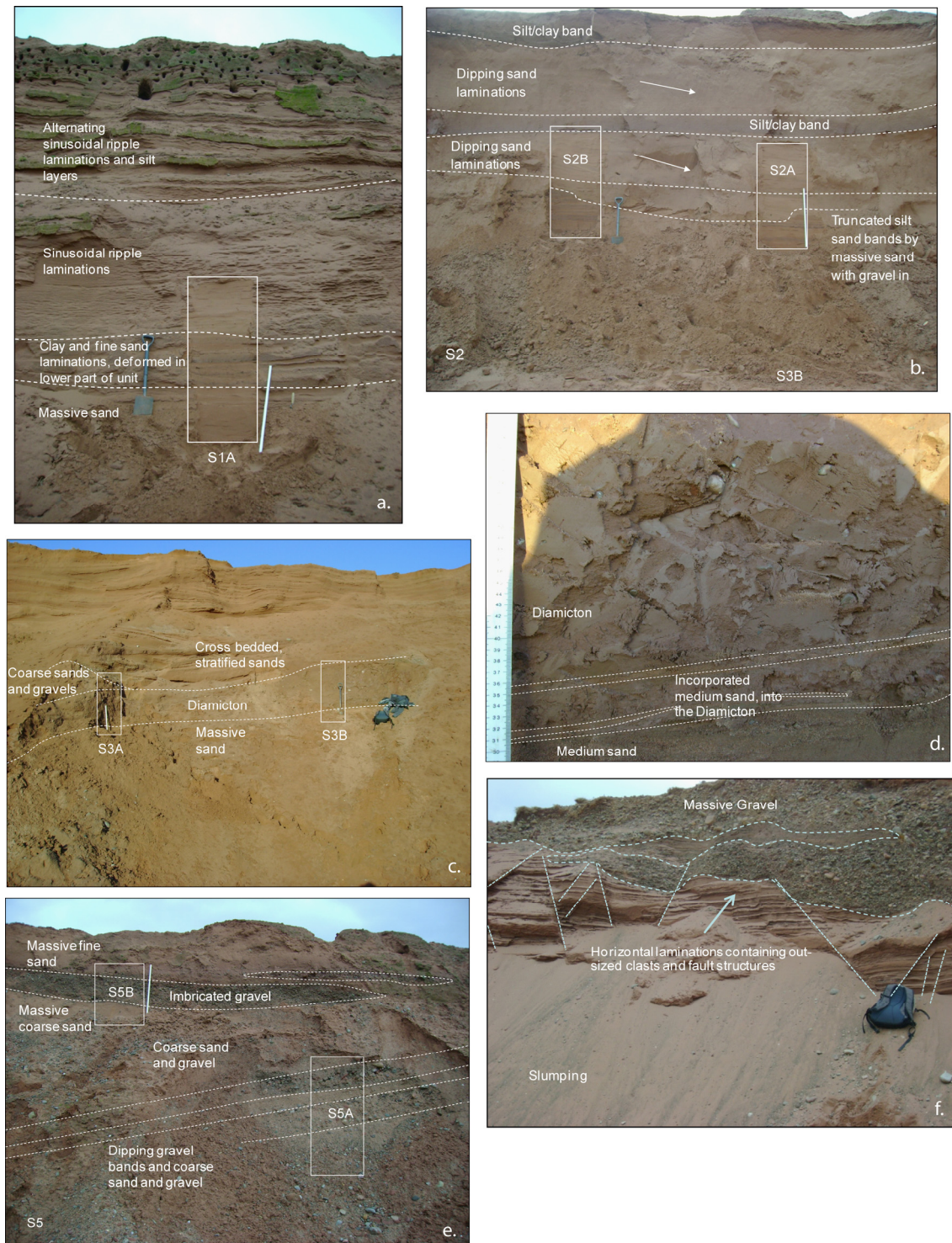


Figure 5.14: Photographs from Faugh sand pit: (a) LFA F1: alternations of fine sand laminations and ripples structures; (b) LFA F1: shallow dipping fine sand and thick fine-grained bands; (c) LFA F3: interdigitated diamicton, massive sand and cross-bedded stratified sand and granule gravel; (d) LFA F3: sheared, incorporated lower boundary of the diamicton; (e) LFA F4: interbedded, clast supported granule-pebble gravel and stratified coarse sand with some gravel, massive coarse sand and wide, thin imbricated gravel lithofacies; and (f) LFA F4: heavily faulted laminated sand and massive gravel lithofacies.

LFA 1 - interpretation

Sediments in LFA F1 are interpreted to have been deposited in a low energy environment. The sinusoidal ripples, clay-silt laminae, clay bands and laminated fine sands are indicative of rapid suspension settling within quiet waters (Jopling & Walker, 1968; Ashley *et al.*, 1982). Upward transitions from sinusoidal to type 'B' ripples are consistent with an increase in bedload transport relative to suspension fall-out (Jopling & Walker, 1968), whilst horizontally bedded and coarser sand indicate a change to low density turbidity currents (Reineck & Singh, 1975; Allen, 1984; Smith & Ashley, 1985). Palaeocurrent data indicate that flow was towards the northwest. Evidence of cross-bedded sand with minor granule gravel at S2 (Fig. 5.13) is typical of foreset aggradation, abandonment and reactivation (Smith & Ashley, 1985). The alternations between facies type exhibited within LFA F1 suggests that the depositional environment was constantly in flux, with a variety of processes, including density underflows, suspension settling, turbidite sedimentation and foreset progradation taking place.

Lithofacies Association 2 (LFA F2; cross-bedded sands with upper fine-grained drapes) - description

Stratigraphically positioned above LFA F1 are a series of shallow-angled, cross-bedded sands that display a variety of orientations, and are bounded by erosional lower contacts. Occasional shallow, metre-scale cut and fill structures composed of sand are also evident throughout the sequence. These facies are overlain by drapes of fine-grained sand and clay bands which exhibit a cyclical series of climbing ripples and planar sand laminations (S4, Fig. 5.13).

LFA 2 - interpretation

LFA F2 is interpreted as the deposit of a low energy, highly variable fluvial system. The cross-bedded sand is characteristic of dune forms (Miall, 1977), with the various orientations of the dunes indicative of transient, unconfined inflows into the system. Vertical accretion of discrete co-sets bounded by re-activation surfaces demonstrates dune migration (Banerjee & McDonald, 1975). The shallow angled cut and fill sequences are symptomatic of higher energy scouring from underflows (Collinson, 1996), whilst the return to fine-grained drapes of laminated fine sand and climbing ripples as seen in LFA F1, suggests that the final stages of deposition were dominated by suspension in standing water and low energy traction currents (Jopling & Walker, 1968; Smith & Ashley, 1985).

Lithofacies Association 3 (LFA F3; sands and gravels with incorporated diamicton) - description

Situated in the north-eastern corner of the flat-topped hill, LFA F3 is dominated by laminated, cross-bedded and massive sand, gravel channels and interbedded diamicton (S3, Fig. 5.14c). The bottom 1 m of the section is composed of laminated sand and coarse, massive sand. In the eastern-most exposures the sand grades up through massive red silt, into a 1 m thick, 'spoon-shaped' channel composed of pebble-cobble gravel and coarse sand, which pinches out at either end (S3D, Fig. 5.13), and fines upwards and outwards away from the scour bottom. Gravel within the channel reaches diameters of up to 20 cm, and is composed almost exclusively of Permo-Triassic sandstone. The bottom contact with the silts is erosional, with evidence for both incorporation and normal faulting. The gravel channel grades up into a *ca.* 2 m thick stratified sand and gravel sequence, dipping towards the west, horizontally aligned granule gravel facies and massive coarse sand (S3C, Fig. 5.13).

West of the gravel lithofacies and stratigraphically positioned above the massive sand lithofacies is a diamicton. The diamicton is 1.3 m thick, massive, with a reddish (2.5YR 3/3), sandy-silty, matrix-dominated texture and a soft to firm composition (Fig. 5.14d). Clasts are rounded to sub-rounded and up to 10 cm in diameter, although generally scarce. Larger clasts are generally positioned towards the middle of the lithofacies, with the upper 40 cm containing 1 - 3 cm (diameter) soft sediment inclusions. The diamicton unit dips down towards the west (towards LFA F1 & 2), pinching out at both ends with an interbedded (sheared) lower contact (Fig. 5.14d). Resting on the diamicton are sand and laminated silt facies (showing evidence of truncation), and a very coarse sand and granule gravel fining upwards into cross-bedded sequences typical of LFA F2 (S3A, Fig. 5.13). The finer sands contain a series of 20 - 30 cm contorted clay-silt structures, characterised by recumbent folds. Balls of diamicton have been incorporated within the sand facies lying directly on top of the interbedded diamicton.

### LFA 3 - interpretation

The series of laminated and massive, coarse sand lithofacies are analogous to deposition by low density turbidity currents (Reineck & Singh, 1975; Allen, 1984; Smith & Ashley, 1985) with the massive sands probably associated with sediment gravity flows (Eyles, *et al.*, 1987). The incorporated contacts and westerly dipping lithofacies support this interpretation (Lawson, 1979, 1981, 1982). The thickness (1.3 m) and geometry of the diamicton coupled with the small shear zone of incorporated and deformed material at the base is indicative of a debris flow (Lawson, 1979, 1982). Soft sediment pods are thought to have formed by the incorporation of ponded subaerial material (Lawson, 1982). The series of massive sand, laminated fine sand and normally graded coarse-fine sand resting on-top of the diamicton are again interpreted as sediment gravity flows associated with turbidity currents, with the recumbent fold structures, convolute bedding and diamicton balls supporting this notion (Lawson, 1981, 1982). The gravel channel, exposed to the east of the diamicton, has a geometry typical of a scour structure, with interdigitated sand and

granule-pebble gravel in the shallow part of the scour and fining upwards typical of a channel bend, characterised by lower energies and point bar growth down the avalanche face in the inner bend, and high energy scour conditions in the thalweg (Miall, 1985). Westwards dipping, tabular cross-beds of coarse sand and granule gravel were probably formed by dune accretion (Miall, 1977, 1985; Collinson, 1996).

#### Lithofacies Association 4 (LFA F4; gravel channels and cross bedded gravel and sand) - description

The eastern corner of the quarry is characterised by interbedded NNE dipping (32°) stratified coarse sand with granule gravel, clast-supported granule gravel and clast-supported pebble gravel (with each lithofacies less than 0.4 m thick), occurrences of clay balls and gradational contacts (S5A, Fig. 5.13). This leads stratigraphically into *ca.* 1 m of massive coarse sand with some granule gravel, followed by a 0.55 m thick pebble-cobble gravel channel fill characterised by an erosional basal contact and gradational fining into dark-red fine sand and silt (S5B, Fig. 5.13). The gravel channel is over 10 m wide, pinching out at either end, with a flat basal contact (Fig. 5.14e), and represents one of many similarly formed gravel facies. Imbrication within the gravel facies is orientated towards the northwest (330°), with cobbles up to 20 cm diameter. The gravel lithofacies contains evidence of normal faulting, with offsets of up to a meter, diamicton balls, and rare blocks of sand *ca.* 30 cm thick. Lying stratigraphically above this sequence, in the very north-east of the quarry, are further massive, pebble-cobble gravel channels containing inter-fingered sand lithofacies and heavily faulted laminated fine sand with occasional outsized granule-gravel clasts (Fig. 5.14f). The faults are normal and downthrow is towards the WSW and ENE at dips of 50-70°. They extend up into the gravel channels and have offsets ranging from centimetre to metre scale (Fig. 5.14f). The massive gravel deposits have erosional lower contacts.

#### LFA 4 - interpretation

The dipping, stratified, coarse sand and granule-pebble gravel is thought to represent lateral accretion of a transverse bar form, 1.5 m thick (Collinson & Thompson, 1989). The beds dip obliquely from the general palaeocurrent direction displayed in the imbricated upper gravel facies, indicating lateral shoreward accretion, which is typical of transverse bars (Boothroyd & Ashley, 1975; Miall, 1977, 1985; Collinson & Thompson, 1989). The cyclical nature pertaining to pulsed changes in flow velocity and depth is elucidated from the facies organisation (Miall, 1985; Collinson & Thompson, 1989). The massive nature of each individual facies could result from rapidly fluctuating discharges during which the sediment concentrations were high (Collinson & Thompson, 1989). Flat, long channel geometries with tapered ends, imbrication and massive gravel beds is demonstrative of longitudinal bar deposition dominated by bed-load transport in a fluid

flow (Miall, 1977, 1985). Gravel imbrication demonstrates traction deposition towards the NW with flow thus corresponding to the orientation of the ridge north of the exposure. Fining upwards sequences suggest gradual abandonment of bars (Miall, 1977), whilst blocks of sediment within the bars probably resulted from bank failure. Coarse grained sand and granule gravel, within which these bars are deposited, is symptomatic of a high energy, non-cohesive and sediment rich fluvial system. The uppermost facies, composed of interdigitated horizontally laminated sand and massive gravel, demonstrates deposition in the upper flow regime. The horizontal sand laminations were deposited during planar flow conditions (Allen, 1984, Collinson, 1996), and the sheet gravels resulted from high competence, rapidly accreting, rising stage flows, possibly as a result of a major flood event (e.g. Russell *et al.*, 2001). As the gravels are interfingered with the sand facies it suggests that these gravel channels originated within a central zone, and then accreted laterally across the braidplain during rising stage flow. The normal faults indicate vertical collapse of the depositional floor (McCarroll & Rijsdijk, 2003), possibly due to the melt-out of underlying dead ice. The orientation of the faults (striking transverse to the general palaeocurrent direction and ridge orientation) is typical of marginal collapse caused by the removal of ice-supporting walls (McDonald & Shilts, 1975).

#### 5.5.2.2 *Whin Hill, Faugh sand and gravel pit (Huddart, 1970)*

Huddart (1970) investigated Faugh sand pit at Whin Hill (NY 515 552), although only the top 6.2 m were exposed when he gained access (Table 5.1).

<i>Dip</i>	<i>Stratigraphy</i>	<i>Thickness (m)</i>	<i>Depth (m)</i>
16°	Horizontally stratified, coarse sand with occasional pebble-gravel	1.7	1.7
16-18°	Horizontally stratified, coarse sand with occasional pebble-gravel	3.0	4.7
2°	Parallel lamination in silts and fine sands with slightly sinusoidal lamination and occasional pebble-gravel	1.5	6.2

Table 5.1: Whin Hill, Faugh (Huddart, 1970)

The exposure displays structures which are typical of a deltaic sequence including fine grained, parallel laminated sand (1.5 m) analogous to bottomsets, and dipping, stratified coarse sand with pebble gravel (3.0 m), which are characteristic of foreset structures (Smith & Ashley, 1985). Imbricate gravel at the site gave orientations of 205° (Huddart, 1970) suggesting that the sequence prograded in a north-easterly direction, similar to that of the linear ridges. Fault structures



identified throughout the sequence indicate vertical collapse of the depositional floor (McCarroll & Rijdsdijk, 2003).

#### 5.5.2.3 Borehole Data:

Borehole data (Jackson, 1979) was obtained from Crews (NY 533 586), Netherton (NY 541 569) and Whin Hill (NY 514 553) (Figs. 5.1 and 5.12). The boreholes reveal a series of silt, fine-coarse sand and sandy gravel facies, with vertical aggradations up to 25 m thick and a general coarsening upwards (Fig. 5.12). At Whin Hill a thin (1.3 m) bed of diamicton is interdigitated between the sands and gravels. Thus deposition is thought to have occurred in a low energy environment (Collinson, 1996), with the presence of diamicton demonstrating the proximity of ice.

#### Synopsis of quarry exposures and borehole data of flat-topped hills: depositional environment

The stratigraphic architecture and characteristics of the lithofacies associations exposed at Faugh sand pit are analogous to the infilling of a former ice-walled lake plain (cf. Smed, 1962; Clayton & Cherry, 1967; Ham & Attig, 1997; Johnson & Clayton, 2003, Clayton *et al.*, 2008). The central zone of the mound (LFA F1 & 2) is characterised by rhythmically bedded fine-grained sediments, drop-stones and a general absence of deformation, typical of bottomsets (Smith & Ashley, 1985; Clayton *et al.*, 2008). These grade up into sand-dominated cut and fill sequences and dune bedforms, representing low energy glaciofluvial flow across the lake floor from a variety of directions during the final stages of infilling. The marginal deposits are generally coarser, with evidence of debris flows composed of diamicton and sand dipping towards the centre of the lake, as typified by other relict lake plains (e.g. Clayton & Cherry, 1967; Clayton *et al.*, 2008). Foreset structures at Whin Hill could represent a small deltaic sequence (Smith & Ashley, 1985; Aitken, 1995) which probably extended into either an ice-walled lake plain (e.g. Clayton *et al.*, 2008) or supraglacial lake (e.g. Mager & Fitzsimons). The coarsening upwards, fine-grained sediments exhibited within the boreholes have a similar sedimentology to the sequences exposed at Whin Hill and Faugh, which suggests a similar origin, although the lack of detailed data on deformation and sedimentological structures does not allow confirmation of an ice-walled origin.

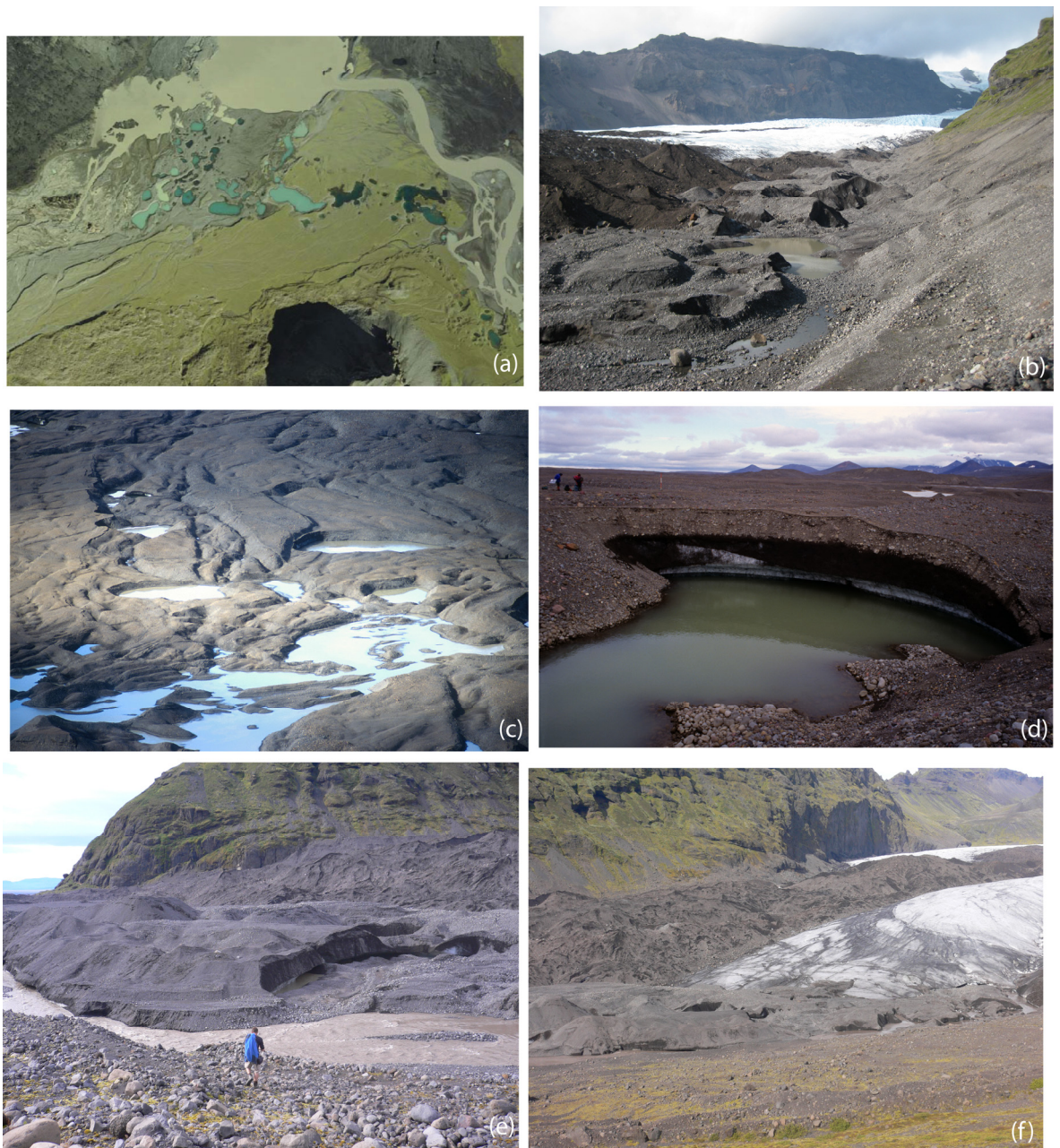
The gravel channels exposed in the north-eastern corner of Faugh sand pit (LFA F4) are envisaged to have been deposited in a broad, supraglacial or ice-walled, braided glaciofluvial trough characterised by migratory bar and dune development, lateral accretion and flood channeling (e.g. Gustavson & Boothroyd, 1987; Huddart, *et al.*, 1999; Russell *et al.*, 2001). The presence of till balls and vertical collapse structures supports this interpretation (Shaw, 1972). Palaeocurrent data indicates that the glaciofluvial channel flowed in a north-easterly direction.



## 5.6 Discussion

### 5.6.1 *Geomorphology and sedimentology*

Geomorphological, stratigraphic and sedimentological analysis of the Brampton kame belt has enabled a number of key inferences to be made about the landsystem. We now provide a synopsis of our interpretations of depressions, ridges, flat-topped hills and meltwater channels based upon the geomorphic and sedimentological details of the field sites described and analysed above, and informed by modern analogues from Iceland (see Fig. 5.15). The actively receding, temperate glaciers of Iceland offer the closest modern analogue to the former margins of mid-latitude Pleistocene ice sheets and as such can provide a wealth of sedimentary and geomorphological information on subglacial processes and landsystems. This synopsis is then used to provide a depositional model for the Brampton kame belt.



< Figure 5.15: Modern Analogues (Iceland): (a) aerial photograph (2007) of Sandsfellsjökull (top-left) and its forefield, revealing a number of lakes; (b) ice-cored eskers and pitted outwash: Sandsfellsjökull; (c) pitted outwash: Bruarjökull; (d) glacier ice revealed beneath pitted outwash: Bruarjökull; (e) outwash covering the snout of Fjalljökull; and (f) ice-cored outwash fan and eskers at Fjalljökull.

Depressions are interpreted as kettle holes, formed by the melt-out of areas of dead ice underlying glacial sediment (e.g. Clayton, 1964; Price, 1971, 1973; Krüger, 1994; Evans & Twigg, 2002). The prevalence of kettle holes and the sedimentological evidence for widespread faulting and subsidence throughout the region is construed as verification of extensive ice cored sedimentary deposition, followed by topographic inversion during ablation. This sort of environmental setting comprising ice-cored glacial outwash is commonly observed along the margins of modern glacier snouts fed by significant volumes of meltwater such as those in Iceland (Evans & Twigg, 2002; Fig. 5.15). Sedimentological and stratigraphic evidence of folding and faulting at both Kirkhouse and Faugh sand pit, combined with observations made by Huddart (1970), demonstrate the presence of vertical collapse structures throughout the lithofacies associations. This supports the notion of either melt-out of underlying dead ice or the removal of ice-supporting walls (Boulton, 1972; McDonald & Shilts, 1975; Johnson & Clayton, 2003). The geomorphology, which displays a series of kettle-hole features within ridges and flat-topped hills, supports this inference.

Ridges contain two discrete sedimentary successions, those at the ridge core and those in the ridge flanks. Ridge cores are generally composed of bars, dunes, channels and scour fills. These structures are typical of braided river deposits exhibiting unconfined flow, variable discharge and channel migration (Boothroyd & Ashley, 1975; Miall, 1977, 1985). Exposures in the ridge margins reveal a lateral fining into a series of dunes, planar bedded sand, ripple structures and fine-grained bands which are analogous to flow in a low energy glaciofluvial environment characterised by fluctuating water depths and periodic abandonment. Extensive faulting both marginally (Brampton ridge, Whin Hill) and pervasively (Faugh) throughout the exposures suggests that ice-contact conditions prevailed (e.g. Price, 1969, 1971, 1973; Shaw, 1972; McDonald & Shilts, 1975). Thus these deposits are analogous to eskers formed in either a broad, supraglacial (e.g. Huddart & Bennett, 1997; Huddart *et al.* 1999; Russell *et al.*, 2001) or/and ice-contact (e.g. Huddart, 1970; Shaw, 1972) trough. Geomorphic evidence for complex dead ice disintegration is manifest in the beaded, discontinuous morphology and pockmarked terrain (cf. Price, 1969, 1971, 1973; Huddart & Bennett, 1997; Thomas *et al.*, 1998; Evans & Twigg, 2002). Despite individual ridge lengths rarely exceeding 1 km, networks of ridges can be traced along the entire length of the Brampton kame belt, thereby demarcating a series of beaded esker-like morphologies (e.g. Banerjee & McDonald, 1975). This undulating ridge morphology is archetypal of ice cored sediment undergoing differential melting, collapse and topographic inversion (Price, 1966, 1969, 1973; Boulton, 1972; Evans, 2009) and can be observed in evolving ice-marginal landsystems along the margins of Sandsfellsjökull, Bruarjökull and Fjalljökull in Iceland (e.g. Evans & Twigg, 2002; Fig.

5.15b,c,e,f). These active margins demonstrate the temporal changes that ice-cored glacial sediments undergo as topographic inversion occurs, leading to the gradual disintegration of the ridge structure, localised ponding and debris flow reworking (e.g. Evans & Twigg, 2002; Fig. 5.15). Marginally deposited debris flows composed of diamicton intercalated between glaciofluvial deposits, supports the interpretation of an ice-walled depositional environment.

The two major ridge orientations, together with associated palaeocurrents suggest that flow generally drained north-eastwards and north-westwards. This indicates that the meltwater drainage network was sourced in the Eden Valley and northern Pennines. There is some evidence for cross-cutting between the two orientated ridge systems, especially visible in the aerial photographs, which, when combined with the buried landforms, suggests that the kame belt formed either time transgressively in a series of stages as ice receded out of the Solway Lowlands or at various levels within the ice (i.e. supraglacially, englacially and subglacially). The Eden Valley/north Pennines provenance for the ridge material is verified by the clast lithological analysis. The diamicton lithofacies at Faugh sand pit is composed of 47% Silurian and Carboniferous sandstone (sourced from the Southern Uplands and Vale of Eden), and then 7% Permo-Triassic sandstone, 4% Carboniferous Limestone, 6.5% Quartzitic sandstone, 6% mudstone and siltstone, 14% Borrowdale Volcanic Group lavas and 0.5% Threlkeld granite sourced from the Vale of Eden and Lake District. The gravel has a similar lithology, although with much less Silurian and Carboniferous sandstone (25%), and more Permo-Triassic sandstone (12%), Carboniferous Limestone (8%) and Borrowdale Volcanic lavas (30%).

Ridge networks are intimately associated with the other major geomorphological features (flat-topped hills and kettle holes) within the kame belt. Some ridges, trending in and out of the flat-topped hills, are partially buried, suggesting that flat-topped hills formed at a slightly later stage of deglaciation, but are related to the same glaciofluvial drainage system.

The Brampton kame belt exhibits a general fining westwards and northwards (Jackson, 1979). This trend can be reconciled with a southerly water source, coming off the Pennine escarpment and Penrith sandstone ridge depositing coarser material proximal to the active ice margin. This interpretation ties in with both the palaeo-current and provenance data. The southern zone of the kame belt is dominated by chaotic hummocky sand and gravel deposits with little structure, thus suggesting that the dominant process was dead ice disintegration and topographic inversion consistent with a relatively thick ice core.

Flat-topped hills are interpreted as ice-walled lake plains, occasionally associated with ice-contact (supraglacial) deltaic sequences (cf. Smed, 1962; Clayton & Cherry, 1967; Johnson & Clayton, 2003; Clayton *et al.*, 2008). The central zone of the flat-topped hill at Faugh is composed of cyclical series of laminated sand, climbing ripples, clay and silt bands and occasional foreset structures, with the upper exposures exhibiting cut and fill, and dune forms. This is typical of a low

energy environment, characterised by fluctuations in the relative importance of suspended sediment rain-out and density underflows (Smith & Ashley, 1985) followed by low energy glaciofluvial accretion across the infilled lake-floor. The absence of rimmed mounds and the thickness of the deposits imply that the ice walled lake plains formed in a stable environment with a large supraglacial debris cover and slow rates of ablation (cf. Clayton *et al.*, 2008). The lack of well developed foreset structures at Faugh sand pit and the variously orientated dune forms suggests that supraglacial drainage, at this location, was both transient and in a state of flux. Modern analogues in Iceland reveal the prevalence of such ice-walled lakes within ice-cored kame terraces and proglacial outwash (e.g. Evans & Twigg, 2002; Fig. 5.15a,b,c,d).

The centres of the flat-topped hills are relatively undisturbed with faulting generally limited to small, centimetre scale faults and folds (e.g. LFA F1 & 2; Fig. 13). This indicates that the ice was perforated almost to the glacier substrate (Dyke & Evans 2003). The marginal zones, in contrast (e.g. LFA F3; Fig. 5.13), are heavily disturbed, as indicated by major debris flow deposits, large scale faulting and interbedded diamicton and stratified units (Eyles, 1979; Eyles *et al.*, 1987; Mager & Fitzsimons, 2007) associated with gravity flows into the lake and vertical collapse following the removal of the ice walls. The geomorphology shows evidence for some small kettle-holed features within the flat-topped hills. This is consistent with an ice-walled model of deposition (e.g. Clayton & Cherry, 1967; Clayton *et al.*, 2008). Flat-topped hills occupy the central tract of the kame belt, often occurring in juxtaposition and linked by series of discontinuous ridges, attesting to their origin as infilled collapsed tunnel systems in a well developed glacier karst (Clayton, 1964).

The kame belt is implicitly associated with a complex meltwater drainage system, situated both to the north and the south of the glaciofluvial complex. Type 1 channels on the Pennine escarpment are orientated predominantly in a SE-NW direction, parallel to the flow of ice up the Vale of Eden during the last glaciation and obliquely to the gradient of the Pennine escarpment (Livingstone *et al.*, 2008). Their anastomosing often complex morphology, length and undulatory long profiles all imply a subglacial formation, whereby topography had little influence on meltwater flow (cf. Greenwood, *et al.* 2007). On the Penrith sandstone ridge Type 1 channels are intimately associated with eskers and ice-contact deltas and have a greater relationship with the topography of the ridge (e.g. the meltwater channels do not show a significant association with ice flow in the region). This suggests that they formed at a late stage of deglaciation, near to the ice margin (cf. Glasser & Sambrook-Smith, 1999). Type 2 channels have a strong topographic influence, orientated parallel to the contours of the slope and perched on the flank of the Vale of Eden and Penrith sandstone ridge. These channels are therefore thought to be ice-marginal channels formed during ice downwasting (Dyke, 1993; Hätterstrand, 1998). Type 3 channels have similar morphologies to 'chute' channels described by Sissons (1960, 1961), Russell (1995) and Glasser & Sambrook-Smith (1999) and are thought to have formed by meltwater being rapidly routed into the subglacial system from the ice margin. This is supported by their association with ice-marginal channels and

their spectacular incised profiles which plunge down gradient either into or towards the subglacial network of the Pennine escarpment (Type 1 channels), or into the present day River Eden. This meltwater system infringes upon the southern-most margin of the kame belt resulting in the dissection of the glaciofluvial deposits into elongate erosional remnants by meltwater channels. The Eden Valley, which is deeply incised into Permo-Triassic sandstone, could potentially have been a tunnel valley, capable of transporting significant amounts of sediment into the kame belt. This tentative conclusion is based on the alignment of many meltwater channels on the Pennine escarpment, which trend towards the current River Eden (Fig. 5.5c), and the south-western arm of the kame belt which trends out of the Eden Valley (Fig. 5.5a). Meltwater also entered the Brampton kame belt from off the Penrith sandstone ridge, with several flights of small channels cutting across the ridge (Livingstone *et al.*, 2008).

A major subglacial meltwater channel sourced from the Hallbankgate esker cuts through the Irthing-South Tyne watershed into the Tyne Valley drainage system in the northern-most zone of the kame belt. This could either have acted as a proglacial over-spill channel from water running through the kame belt and therefore a time transgressive relict of subglacial flow across the watershed, or it could relate to active subglacial flow across the watershed as part of a major ice drainage network. Its undulatory nature suggests the latter (Paterson, 1994).

#### 5.6.2 *Model of formation for the Brampton kame belt*

The Brampton kame belt is envisaged to have formed in an ice-contact meltwater drainage network that evolved spatially and temporally from a sub-marginal and subglacial system to a progressively supraglacial system through the enlargement of a complex glacier karst (*sensu* Clayton, 1964) (Fig. 5.16).

The presence of an extensive subglacial meltwater system leading into (Pennine escarpment & Eden Valley) and out of (Hallbankgate meltwater channel) the kame belt suggests that initial drainage was subglacial, as part of a combined Vale of Eden-Tyne Gap drainage network (LT1-4: Chapter 2) (Livingstone *et al.*, in press a). However, as ice from both Scotland and the Lake District retreated westwards back across the Tyne Gap prior too, during and following the 'Blackhall Wood re-advance' (LT5: Chapter 2) (Trotter, 1929, Livingstone, *et al.*, in press a, b) ice downwasted and stagnated within the lee of the Pennine escarpment and Penrith sandstone ridge (rising up to 260 m O.D.), both of which probably acted as pinning points (e.g. Ó Cofaigh *et al.*, 1999). Hence, initial deposition of the kame belt likely involved subglacial esker development (Brennand, 2000) (Fig. 5.16a: stage I). Esker sedimentation would have been driven by confluent meltwater, the presence of abundant sediment eroded from the Permo-Triassic sandstone and compressive flow and stagnation of ice against the reverse slope of the Tyne Gap (cf. Punkari,



1997, Mäkinen, 2003). Some of the higher and longer eskers, which often lead into the meltwater channels, such as the Hallbankgate esker, Brampton ridge and the S-N orientated esker trending out of the Eden Valley (Fig. 5.5a, b), are likely examples of these subglacial features. Cyclical sedimentation and fining outwards corresponds to a variable discharge regime, with the marginal-subglacial meltwater associations along the Pennine escarpment able to facilitate rapid (seasonal) changes of meltwater flow into the subglacial network (Sissons, 1961; Brennand & Shaw, 1996; Mäkinen, 2003).

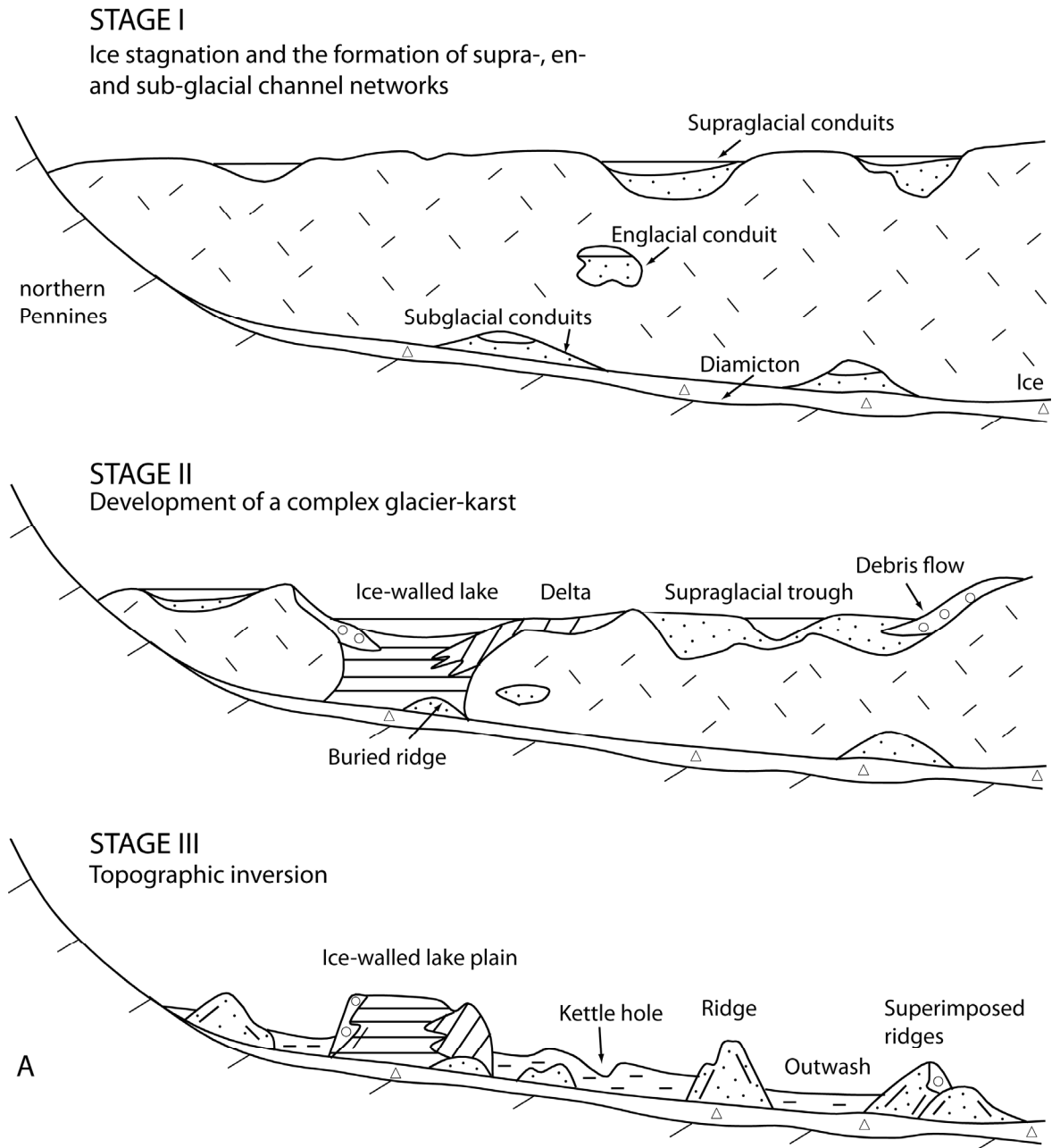
As the ice continued to stagnate and downwaste the early phase of esker sedimentation evolved into a complex 'glaciofluvial moraine' similar to many interlobate areas (e.g. Warren & Ashley, 1994; Brennand & Shaw, 1996; Thomas & Montague, 1997; Punkari, 1997; Mäkinen, 2003; Russell *et al.*, 2003; Sharp *et al.*, 2007). This evolution led to the progressive development of glacial karst (Fig. 5.16a: stage II, b), with perforation of ice occurring by strain softening, tunnel collapse, crevasse formation, and supra/englacial meltwater drainage (e.g. Clayton, 1964).

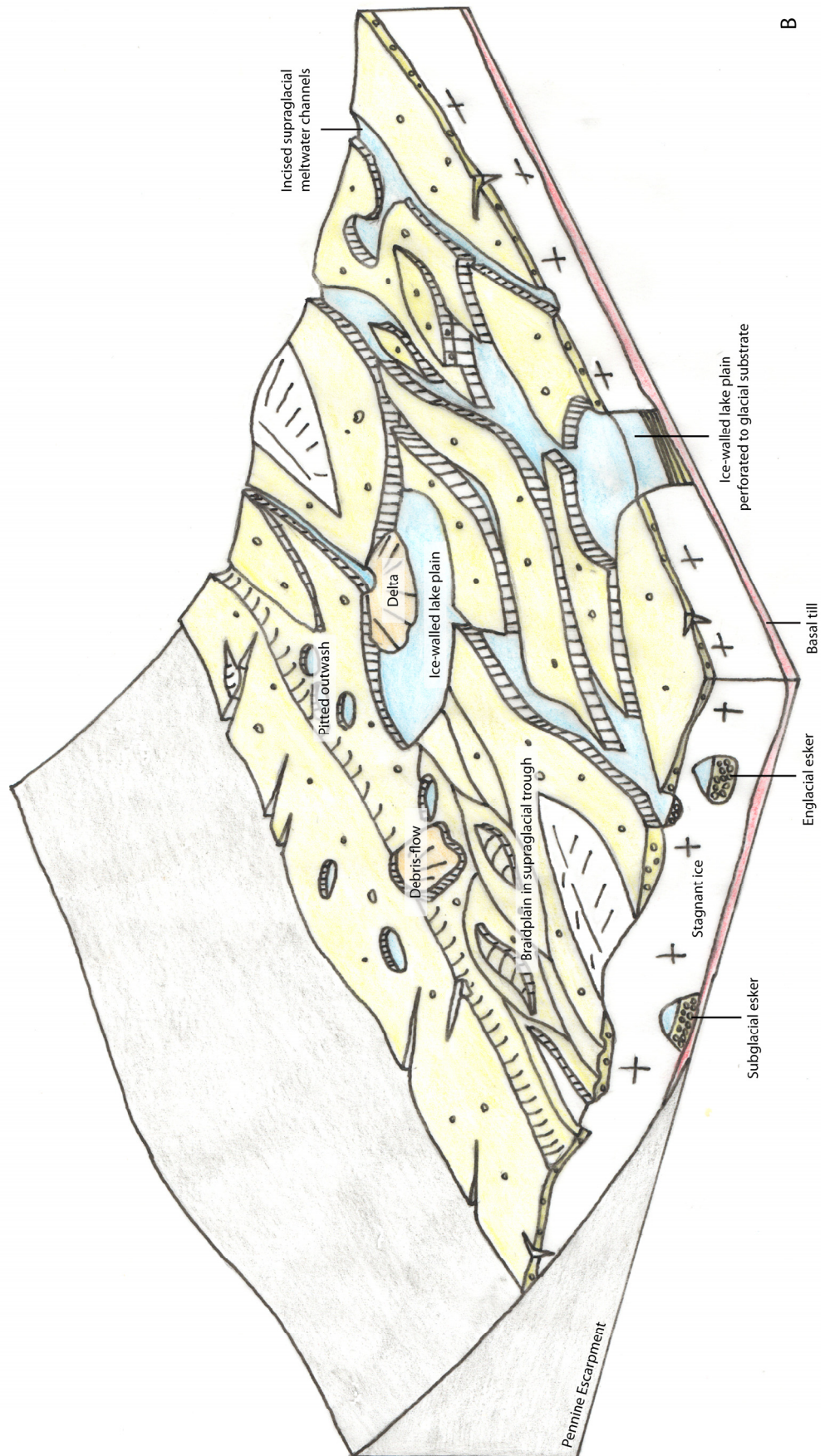
Open channel, supraglacial streams (Fig. 5.16a: stage I/II, b) flowing along the ice surface and constrained laterally by ice walls (e.g. Warren & Ashley, 1994; Bennett & Glasser, 1996) underwent successive phases of capture and abandonment (cf. Thomas & Montague, 1997), resulting in the development of complex braided networks of subdued, discontinuous ridges (Price, 1966, 1969, 1973; Gustavson & Boothroyd, 1987; Huddart, 1999). Sediment architecture reveals glaciofluvial deposits typical of modern braided stream deposits, (e.g. Miall, 1977, 1985) thus attesting to this model of formation. Englacial sedimentation and the development of sub-aerial ice-walled conduits following tunnel roof collapse are also likely to have developed as the ice downwasted (Huddart, 1999) (Fig. 5.16a: stage II, b). As successive stages of topographic inversion occurred multiple changes in drainage configuration could have led to the eventual development of a broad, shallow supraglacial trough as observed at Veggreen, Svalbard (Huddart, 1999) thus accounting for the extensive sweep of continuously deposited glaciofluvial deposits (Fig. 5.16a: stage II, b).

The genesis of ice-walled lakes occurred at a late stage in the development of the kame belt (Fig. 5.16a: stage II, b). The lack of faults within the central portion of the flat-topped hill and the fact that they often overlaid ridges supports the notion that the lakes were perforated right down to the glacial substrate (Dyke & Evans, 2003). Therefore, the kame belt was thought to be initially dominated by subglacial and then englacial-supraglacial meltwater drainage networks which probably exited via the Hallbankgate meltwater channel, thus preventing the development of a lake against the reverse slope of the Tyne Gap, as is typical of many esker complexes (e.g. Banerjee & McDonald, 1975; Gorrell & Shaw, 1991; Brennand, 1994; Mäkinen, 2003). As the ice continued to downwaste and stagnate a thin, chaotic, dead ice-cored apron developed, which prevented the effective discharge of meltwater, therefore causing ponding either in collapsed subglacial tunnels



(Clayton, 1964), or via the expansion of supraglacial lakes into ice-walled lakes due to basal melting (Smed, 1962; Johnson & Clayton, 2003; Clayton *et al.*, 2008). During this stage topographic inversion resulted in the extensive reworking of sediments by debris flows, water escape and vertical collapse. This late-stage chaotic model of supraglacial outwash deposition, lake development, sediment re-distribution and meltwater drainage is epitomised by the multitude of palaeo-currents exhibited within the upper-most sediments (Fig. 5.16b).





< Figure 5.16: (a) Model of formation of the Brampton kame belt (based on the model by Brodzikowski & van Loon, 1991); and (b) schematic diagram of the Brampton kame belt: glacial-karst landsystem.

These final stages in the evolution of the kame belt would have likely involved glaciofluvial supraglacial outwash extending out across the dead-ice cored apron (Rich 1943; Price 1969; Thomas et al., 1985; Gustavson & Boothroyd, 1987; Evans, 2009), leading to the development of hummocky, chaotic, kettle-holed terrain (e.g. Cook, 1946; Clayton, 1964) (Figs. 5.15 and 5.16a: stage III). Spatially this is consistent with the topography of the kame belt, whereby the northern and central zones exhibit striking linearity characterised by the partial preservation of the drainage network despite topographic inversion (Price, 1969; Evans & Twigg, 2002). Conversely, the south-eastern corner exhibits a chaotic kame and kettle landscape, with little evidence of linearity. This is more akin to a thick, extensive deposition of supraglacial ‘pitted’ outwash during a later stage of evolution, and on-top of relatively thick ice (Rich 1943; Price 1969; Thomas et al., 1985; Gustavson & Boothroyd, 1987).

The lack of well developed meltwater channels within the kame belt itself indicates that ice-marginal and proglacial meltwater drainage must have shifted direction during the latter stages of ice recession. South of the Penrith sandstone ridge the Pennine escarpment meltwater network exhibits a time-transgressive morphology, with channels running parallel to each other at successively lower elevations and trending towards the Eden Valley. Thus at some point during deglaciation as the active margin stagnated within the Penrith sandstone ridge (Trotter, 1929; Livingstone *et al.*, 2008) the Pennine escarpment meltwater input into the kame belt must have been severed. The independence of the Penrith sandstone ridge meltwater channels from the final flow phases recorded in the region (Livingstone, *et al.*, 2008), the lower inception points of all channel types and the drainage of a dendritic network of meltwater channels into an ice-contact delta at ~120 m O.D. (Baronwood ice-contact delta – Huddart, 1970) implies formation at a later stage of deglaciation than much of the Pennine escarpment meltwater network.

## 5.7 Conclusions

The Brampton kame belt demonstrates a complex mode of deglacial deposition involving both esker and kame forms/processes (Fig. 5.16) caused by ice stagnating and downwasting in the lee of the Penrith sandstone ridge and north Pennine escarpment. It is composed of flat-topped hills, ridges and depressions, interpreted as ice-walled lake plains, ice-contact meltwater drainage networks and kettle holes respectively. Sedimentation evolved both spatially and temporally from a sub-marginal and sub-glacial drainage system through the enlargement of a complex glacio-karst. This led to the development of englacial and supraglacial drainage systems within the ice-cored terrain, with ponding leading to the development of ice-walled lakes. Topographic inversion led to

the distinctive suite of landforms and represents an end product of a depositional landscape similar to those currently forming at many modern glacier margins (Fig. 5.15). The Brampton kame belt formed during deglaciation, when ice, which was previously converging on the Tyne Gap from Scotland and the Lake District, began to retreat westwards back across the Irthing-Tyne watershed, prior too, during and following the 'Blackhall Wood re-advance' (Livingstone *et al.*, in press b). Further retreat led to the formation of Lake Carlisle and the stagnation of ice within the Penrith sandstone ridge. The meltwater channels along the Pennine escarpment are consistent with the NW direction of the youngest subglacial lineations in the region and therefore record the SE recession of the final ice to occupy the Vale of Eden (LT6: Chapter 2).

Based on the model presented in this paper (Fig. 5.16) a number of general conclusions can be drawn regarding the formation of kame belts:

1. We propose that kame belts demonstrate not only a polygenetic topography (e.g. ridges, flat-topped hills and depressions) but also a time-transgressive evolution with sedimentation controlled primarily by an enlarging glacier karst. The Brampton kame belt demonstrates the potential importance of dead-ice development and an evolving glacier karst in the formation of glaciofluvial depo-centres as it provides a mechanism for both triggering and promoting extensive sedimentation within the ice in various ice-walled settings.
2. Kame formation is shown to be very dynamic, with downwasting, debris flows, melt-out of dead-ice and enlargement of the glacier karst all leading to a shifting glaciofluvial network characterised by both constrained and unconstrained flow, en- sub- and supra-glacial drainage, and ponding. This dynamism explains the complex polyphase and polygenetic geomorphology and sedimentology exhibited within kame belts.
3. By their very nature kame belts are associated with meltwater drainage of the ice mass and therefore should not be treated in isolation of other glaciofluvial and glaciolacustrine landforms with which they are inextricably linked in a spatial and temporal continuum.

## 5.8 Acknowledgments

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## Chapter 6

### **Sedimentary evidence for a major glacial oscillation and pro-glacial lake formation in the Solway Lowlands (Cumbria, UK) during Late Devensian deglaciation**

Livingstone, S.J., Ó Cofaigh, C., Evans, D.J.A. & Palmer, A. In press. Sedimentary evidence for a major glacial oscillation and pro-glacial lake formation in the Solway Lowlands (Cumbria, UK) during Late Devensian deglaciation. *Boreas*.

#### **Abstract**

This paper is a sedimentological investigation of Late Devensian glacial deposits from the Solway Lowlands, NE England, in the central sector of the last British-Irish Ice Sheet. In this region laminated glaciolacustrine sediments occur, sandwiched between diamicton lithofacies interpreted as subglacial tills. At one location the laminated sediments are interpreted as varves, and indicate the former presence of a proglacial lake. Correlation of these varves with other laminated sediment indicate that the glacial-lake was at least 140 km<sup>2</sup> in area and probably much larger. Extensive beds of sand, silt and gravel throughout the Solway Basin associated with the lake demonstrate ice free conditions over a large area. Based on the number of varves present, the lake was in existence for at least 261 years. The stratigraphic sequence of varves bracketed by tills implies a major glacial oscillation in this region prior to the Scottish re-advance (16.8 cal. ka BP). This oscillation is tentatively correlated with the Gosforth Oscillation at ~19.5 cal. ka BP. Subsequent overriding of these glaciolacustrine sediments during a westwards moving readvance demonstrates rapid ice loss and then gain within the Solway Lowlands from ice-dispersal centres in the Lake District, Pennines and Southern Uplands. It is speculated that the influence of this and other lakes along the north-eastern edge of the Irish Sea Basin would have had an impact on ice-sheet dynamics.

#### **6.1 Introduction**

Recent evidence suggests that the Late Devensian (Marine Isotope Stage 2) British-Irish Ice Sheet (BIIS) was characterised by multiple ice-dispersal centres drained by ice streams and fast-flowing outlet glaciers (e.g. Merritt *et al.*, 1995; Boulton & Hagdorn, 2006; Bradwell *et al.*, 2007; Ó Cofaigh & Evans, 2007; Roberts *et al.*, 2007; Hubbard *et al.*, 2009; van Landeghem *et al.*, 2009). Perturbations to the ice-sheet system from both external and internal forcings are therefore envisaged to have been highly sensitive, with enhanced ice-flow conditions rapidly transmitting changes far into the ice-sheet interior leading to relatively rapid, non-linear and complex re-

configurations of the BIIS (cf. Bamber *et al.*, 2007; Evans *et al.*, 2009; Hubbard *et al.*, 2009). This is indicated by the geomorphic and sedimentological evidence, which advocates a dynamic ice sheet characterised by multiple ice-flow switches (e.g. Livingstone *et al.*, 2008; Greenwood & Clark, 2008; Evans *et al.*, 2009) and an oscillatory margin (Eyles & McCabe, 1989; McCabe, 1996; Evans & Ó Cofaigh, 2003; Thomas *et al.*, 2004; Thomas & Chiverrell, 2007; McCabe *et al.*, 2005, 2007). This sensitivity has, in some instances, been correlated to abrupt millennial-scale climate changes occurring in the North Atlantic Ocean (e.g. McCabe & Clark, 1998; Scourse *et al.*, 2000), with recent modelling studies suggesting a dynamic BIIS modulated and phase lagged by external climatic forcings (Hubbard, *et al.*, 2009).

The Irish-Celtic Sea was thought to contain an extensive Irish Sea Ice Stream that advanced to the Scilly Isles at some point during the Main Late Devensian glaciation (Scourse & Furze, 2001; Hiemstra *et al.*, 2006). The Irish Sea Ice Stream is envisaged to have been highly responsive during deglaciation with evidence for multiple marginal oscillations (e.g. Evans & Ó Cofaigh 2003; Thomas *et al.*, 2004; Thomas & Chiverrell, 2007). Re-advances have been identified in northeast Ireland between 18.3 – 17.0 cal. ka BP and after 17.0 cal. ka BP (McCabe & Clark, 1998; McCabe *et al.*, 2005, 2007). This latter readvance has been correlated to Heinrich Event 1 as part of an extensive Irish Sea glacial-response to North Atlantic climate change (McCabe & Clark, 1998). Less chronologically well constrained re-advances include the Gosforth oscillation at ~ 19.5 cal ka BP (Merritt & Auton, 2000) and the Scottish re-advance at ~ 16.8 cal ka BP (Trotter, 1929; Merritt & Auton, 2000) along the Cumbrian coast.

The Solway Lowlands, in the north-east sector of the Irish Sea Basin, (Fig. 6.1) are situated within a palimpsest glacial landscape characterised by multiple phases of flow and containing evidence of a Scottish re-advance (cf. Livingstone *et al.*, 2008). This complexity is partially explained by its location, situated between major ice-dispersal centres of the Pennines, Lake District and Southern Uplands, with ice coalescing and subsequently draining through outlets such as the Tyne and Stainmore gaps (Trotter, 1929; Hollingworth, 1931; Livingstone *et al.*, 2008). The Solway Lowlands also acted as a tributary to the Irish Sea Ice Stream at some point during the Main Late Devensian (Roberts *et al.*, 2007; Livingstone *et al.*, 2008). Stratigraphic and geomorphological evidence suggests that Scottish ice re-advanced into the Solway Lowlands following partial deglaciation, although there is some debate as to the extent of this re-advance (Trotter, 1929; Trotter & Hollingworth, 1932; Huddart, 1970, 1971, 1991, 1994).

This geomorphic complexity manifests itself within the stratigraphic record (see Table 6.1). The oldest glacial deposits associated with the Main Late Devensian Glaciation in the Vale of Eden, Solway Lowlands and Dumfries-shire are the Gillcambon and Chapelknowe Till Formations (Stone *et al.*, in press). These till units have been assigned to the ‘Early Scottish’ advance of ice up the Vale of Eden and across the Stainmore Gap (cf. Huddart & Glasser 2002; Stone *et al.*, in press). An

upper till in the Vale of Eden (Greystoke Till Formation: Stone *et al.*, in press), separated from the Gillcambon Till Formation by sporadic sand, silt and clay deposits ('Middle sands'), is assigned to the 'Main Glaciation' (Dimlington Stadial) (Hollingworth 1931). Goodchild (1875, 1887) and Huddart (1970) attributed all the deposits (including the laminated sediments at Blackhall Wood in Huddart's case) to the subglacial melting of a single, stagnant ice mass, while Hollingworth (1931) surmised that the 'Middle' sands and laminated clays were formed proglacially, thus delimiting a period of partial deglaciation, followed by re-advance. Resolving the uncertainty associated with this glacial re-organisation has major implications for the glacial history in Cumbria in terms of its ice-flow phases, ice-sheet behaviour, glacial-lake development and correlations between regional ice dynamics and the rest of the BIIS.

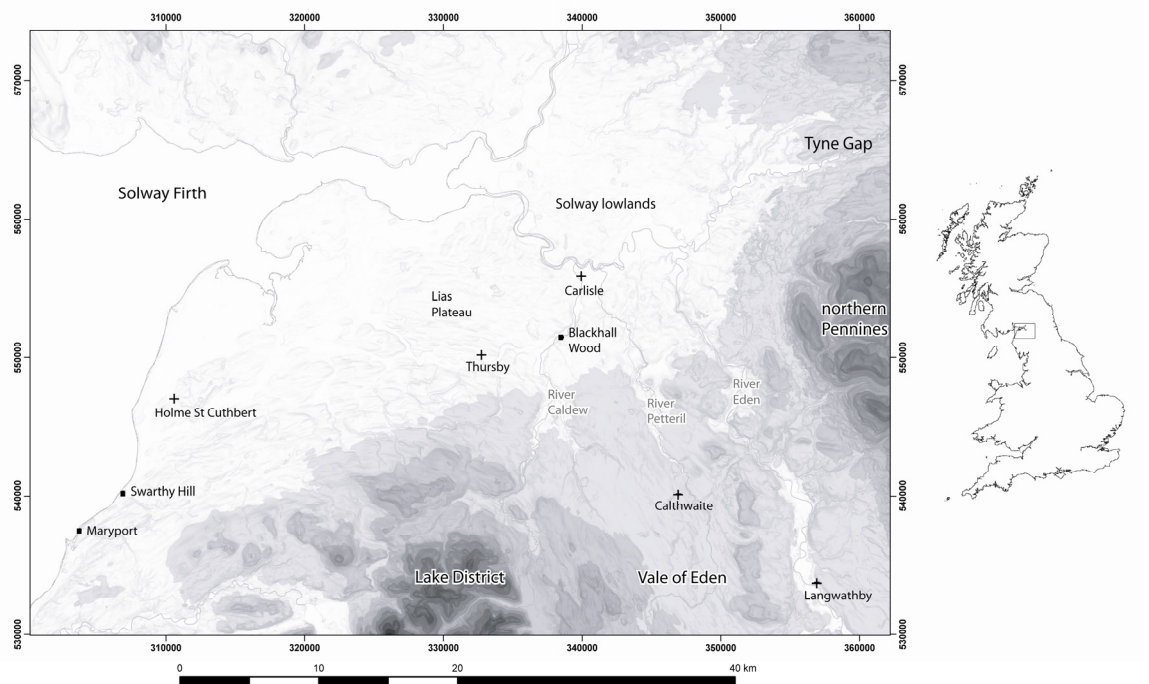


Figure 6.1: Topographic map showing key locations and fieldsites.

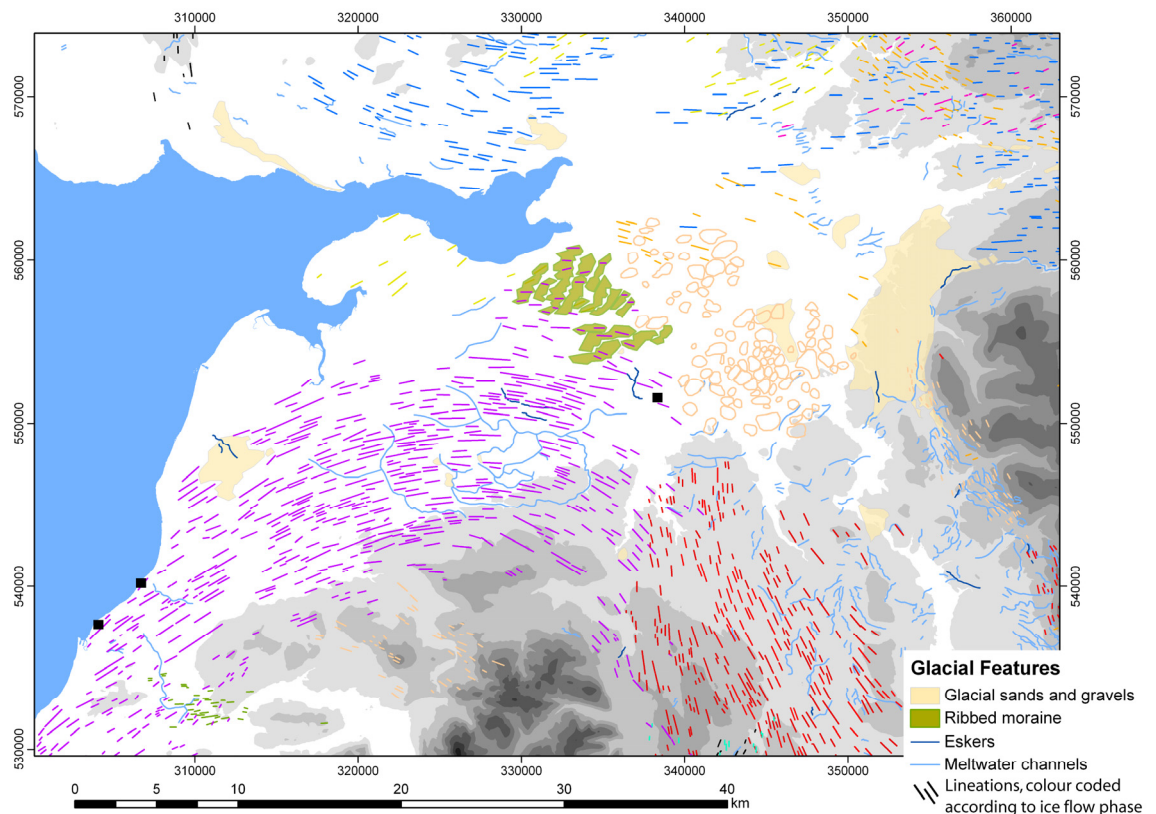
In the northern Solway Lowlands and Dumfries-shire a tripartite division of sediments is recognised, with the Chapelknowe and Greystoke Till Formations overlain by a series of sands and clays (Plumpe Farm Sand and Gravel Formation & Great Easby Clay Formation) and then capped by a thin (< 5 m) upper red till (Gretna Till Formation) (Stone *et al.*, in press). The Great Easby Clay Formation was deposited during formation of Glacial-Lake Carlisle during deglaciation of the region; while the Gretna Till Formation corresponds to a late-stage re-advance of Scottish ice into the Solway Lowlands (e.g. Trotter & Hollingworth 1932). Within the Solway Lowlands the Great Easby Clay Formation has been extended south of Carlisle (Trotter 1929; Huddart 1970), with the tripartite divisions exposed here thought to correspond to a period of deglaciation followed by re-advance of the Scottish ice (Trotter, 1929). However, the Gretna Till Formation underlies drumlins

associated with the arcuate flow of ice around the northern margin of the Lake District and out into the Irish Sea Basin (Fig. 6.2).

This study aims to resolve some of the issues arising from the complex glacial stratigraphy that characterises the Solway Lowlands. Sedimentological investigations of three field sites combined with borehole data and collation of previous field evidence will be used to assess: (i) whether the glacial stratigraphic complexities of the region represent multiple re-advance phases or melt-out of a single stagnant ice mass; (ii) the nature and extent of glacial-lake formation in the region; and (iii) the impact on ice-flow phasing both in the Solway Lowlands and in the northern sector of the Irish Sea Basin.

Event	BGS Lithostratigraphic Formations		Regional Correlation	Date (cal ka BP)
	Cumbria	Dumfries-shire		
Scottish re-advance	Gretna Till Formation		Killard Point Stadial <sup>1</sup>	~ 16.8
Deglaciation (formation of Glacial-Lake Carlisle)	Great Easby Clay Formation	Plumpe Sand and Gravel Formation		
Main Glaciation	Greystoke Till Formation	Chapelknowe Till Formation	Gosforth Oscillation <sup>2</sup>	~ 19.5
Middle Sands			Dimlington stadial	Main Late Devensian
Early Scottish advance	Gillcambon Till Formation			

Table 6.1: Simple Event Stratigraphy for the field area (based on Trotter, 1929; Hollingworth, 1931; Huddart, 1970). Lithostratigraphic formations as used by the British Geological Society (Stone *et al.* in press). Regional correlations from: <sup>1</sup> McCabe *et al.* 1998; <sup>2</sup> Merritt & Auton (2000).



< Figure 6.2: Glacial geomorphology of the Solway Lowlands (Livingstone *et al.*, 2008). Black boxes indicate field sites. The lineations have been colour-coded according to discrete ice-flow phases, with the purple flow-set showing arcuate flow out into the Irish Sea (LT5: Chapter 2), while the red flow-set illustrates the final flow of ice down the Vale of Eden (LT6: Chapter 2).

## 6.2 Study area and glacial geomorphology

The Solway Lowlands (Fig. 6.1) form a low lying (<100 m O.D.) basin bounded by the Solway Firth and Irish Sea to the west and upland terrain to the north, south and east. Two mountain passes, the Vale of Eden and Tyne Gap which run SE-NW and W-E respectively, converge on the Solway Lowlands (Fig. 6.1). The underlying bedrock geology is Permo-Triassic sandstone and mudstone (Hollingworth, 1931).

The area is predominantly composed of Late Devensian deposits that show evidence for extensive glacial activity (Fig. 6.2), as recorded by a dense coverage of subglacial lineations, meltwater channels, ribbed moraine, hummocky terrain, eskers and glaciofluvial accumulations, primarily below 400 m O. D. (Livingstone *et al.*, 2008). The glacial geomorphology has been used to reconstruct a relative chronology of ice-flow phases based on cross-cutting relationships (Chapter 2, Fig. 2.5; Livingstone *et al.*, 2008; Fig. 6.2). Erratic trains suggest that the earliest phase of flow was southwards from Scotland, up the Vale of Eden and across the Stainmore Gap ('Early Scottish advance' of Hollingworth, 1931). This phase is not recorded within the geomorphic record, suggesting that the evidence for it had been destroyed by subsequent ice flows. The next phase to influence the Solway Lowlands was characterised by convergent flow, sourced from Lake District and Southern Upland ice-dispersal centres, moving eastwards into and through the Tyne Gap (LT1-3: Chapter 2) (Livingstone *et al.*, 2008, Fig. 6.2). A major flow switch occurred when ice drained westwards into the Irish Sea Ice Basin (LT5: Chapter 2), probably as a trunk zone feeding into the Irish Sea Ice Stream (Roberts *et al.*, 2007). Superimposed on top of this flow-set is a deltaic complex located at Holme St Cuthbert (Huddart, 1970; Huddart and Glasser, 2002) demarcating the re-advance of ice across the Solway Firth from Scotland (SF1: Chapter 2) (cf. Salt & Evans, 2004). The final ice flow phases are typified by topographically constrained flow lobes extending into the lowland region (LT7: Chapter 2) (Livingstone *et al.*, 2008, Fig. 6.2), which probably occurred prior to, or concurrently with, the Scottish re-advance.

## 6.3 Methods

Field exposures at Blackhall Wood, Maryport and Swarthy Hill (Fig. 6.1) enabled detailed stratigraphic logging and sedimentological analysis to be carried out. These field sites are located

in lineations that are associated with the movement of ice into the Irish Sea Ice Basin (LT5: Chapter 2; Fig. 6.2). Texture, sedimentary structure, colour (Munsell colour chart), bed geometry, contacts and inclusions were all measured and logged, from which lithofacies were identified. Lithofacies codes are based upon those of Evans & Benn (2004). Scaled section sketches were drawn at the larger exposures so that the lateral extent of the lithofacies could be assessed. Clast macrofabric analysis of the a-axis orientation and dip, and investigations of clast lithology supplement the geomorphological mapping of ice-flow pathways. Diamicton samples for thin-section analysis were collected to provide detailed micromorphological information. Borehole logs (British Geological Survey) provided a less detailed but wider coverage, allowing regional stratigraphic correlations. At Blackhall Wood, overlapping monoliths were collected from a 1.3 m sequence of laminated clay and silt, and a continuous set of thin sections was produced for the sequence. Standard techniques were used for the impregnation and preparation of thin sections of unconsolidated sediments at the Centre for Micromorphology, University of London (Palmer *et al.*, 2008a).

## 6.4 Sedimentology and stratigraphy

### 6.4.1 *Blackhall Wood:*

Blackhall Wood (NY 386 515) is situated 5 km south of Carlisle on the banks of the River Caldew (Fig. 6.1). The section is exposed in the north-western edge of a SE-NW trending drumlin (40 m O. D.) that forms part of an arcuate flowset sweeping around the northern edge of the Lake District (LT5: Chapter 2; Fig. 6.2). Late Devensian sediments rest on thinly bedded Permo-Triassic sandstones which dip steeply towards the east. The sediment lithofacies and lithofacies associations are depicted diagrammatically in Fig. 6.3 and described below.

#### Lithofacies 1 and 2: lowermost diamictons:

##### Description:

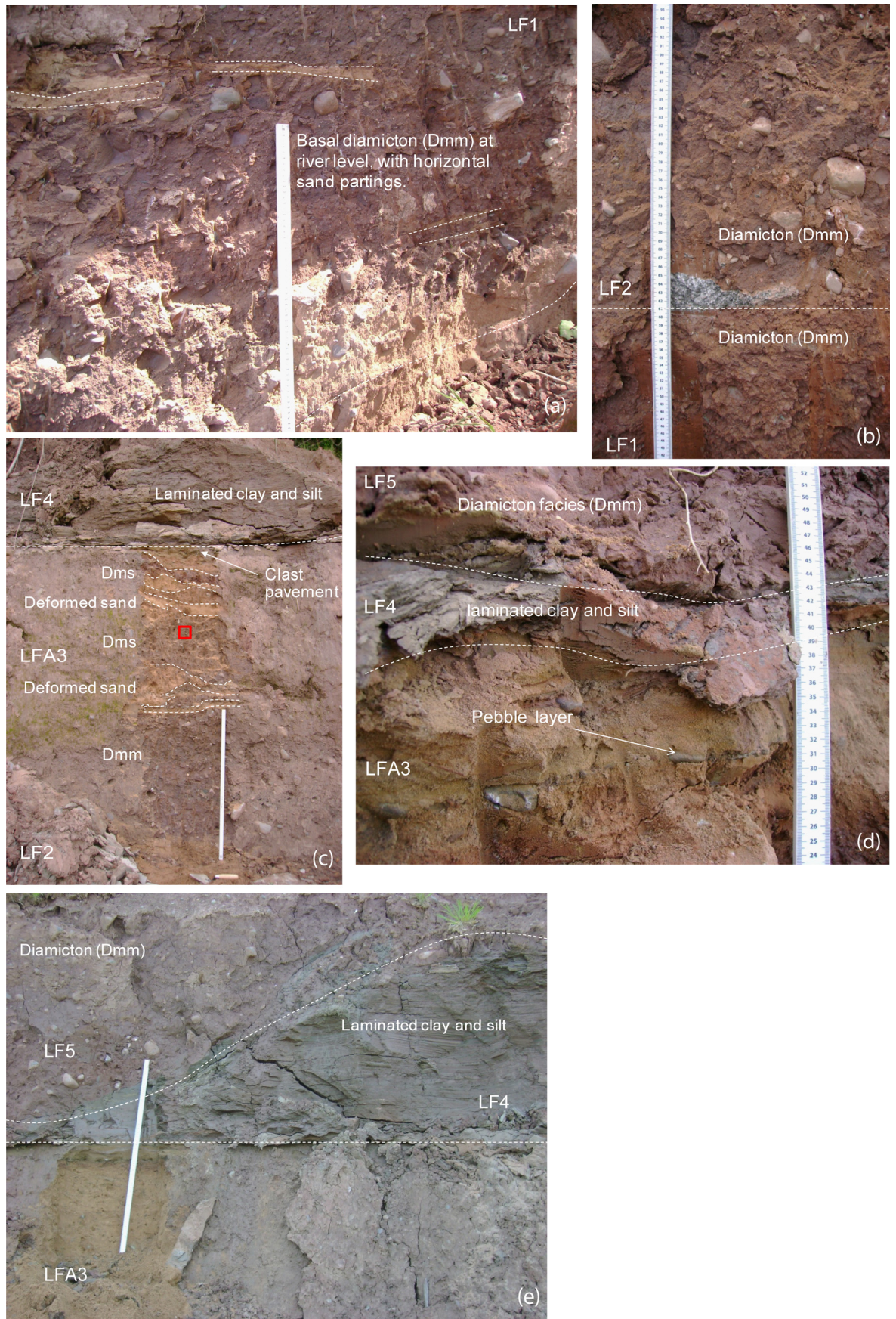
The basal lithofacies consists of a brownish-black (7.5 YR 3/2) matrix-supported diamicton (LF1) up to 2.0 m thick (Figs. 6.3 and 6.4a). The diamicton contains a moderate distribution of predominantly sub-angular to sub-rounded clasts many of which are striated. The principle lithologies (Table 6.2) are Silurian greywacke (31%), Carboniferous limestone (15%), Permo-Triassic sandstone (9%), sandstone (13%) and dolerite (8%), with Southern Upland granite (2.0%) and lavas of the Borrowdale Volcanic Group (0.8%) also identifiable. Horizontal (sometimes wavy), millimetre thick, sand partings occur throughout the lithofacies (Fig. 6.3). A thin section taken near the top of the lithofacies and orientated S-N exhibits lineations, turbate structures and an interconnected network of sub-horizontal and sub-vertical fissures (Fig. 6.5a). This fissility is also



noted macroscopically (Fig. 6.4a, b). Two fabrics (Fig. 6.6) indicate a WSW–ENE orientation with  $S_1$  eigenvalues of 0.61 and 0.55.

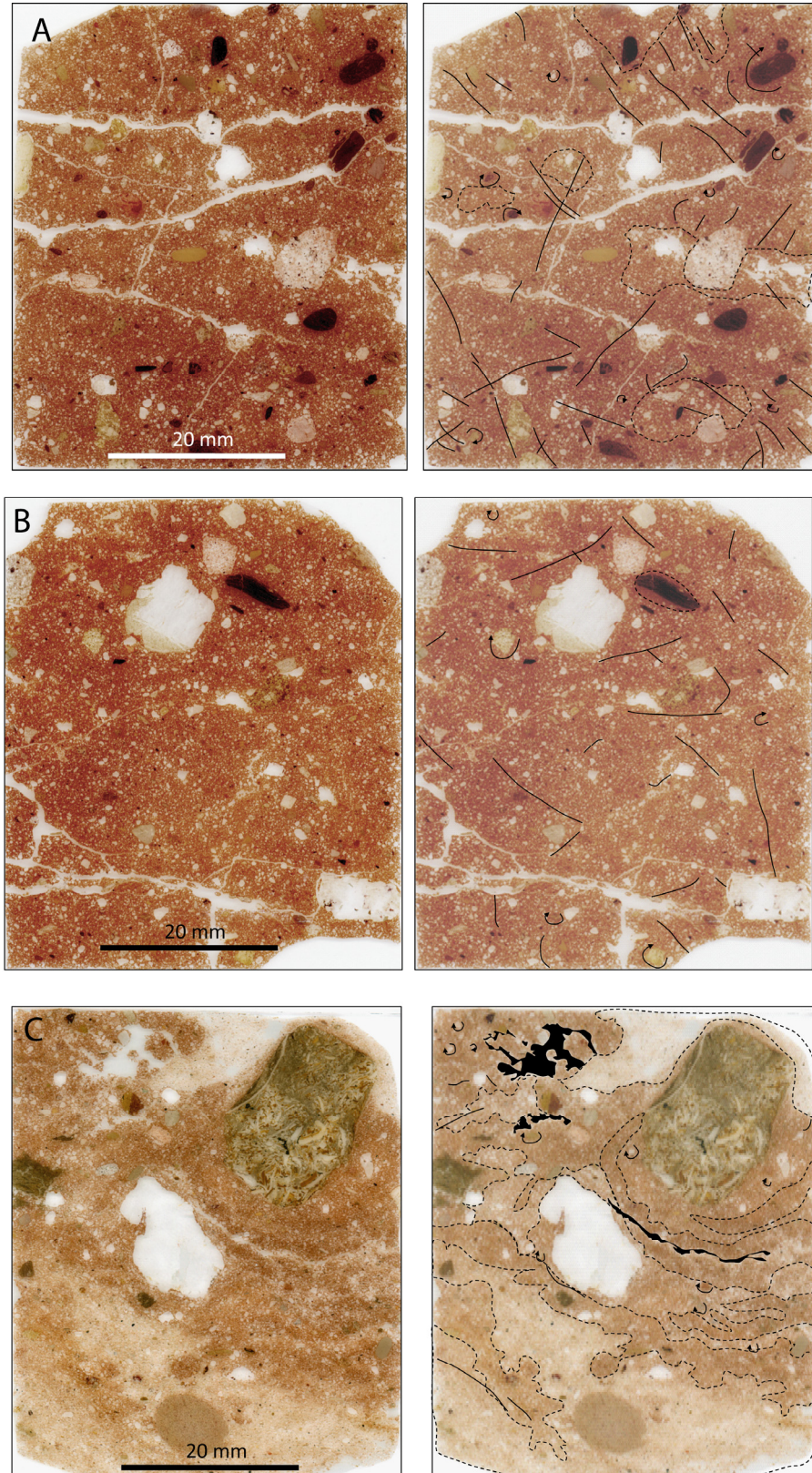


< Figure 6.3: Blackhall Wood section logs from a ~ 20 m long cliff exposure. Thin section letters refer to Figure 6.5.





< Figure 6.4: Pictures of Blackhall Wood lithofacies: (a) LF1 characterised by sand partings and a high degree of fissility; (b) indistinct boundary between LF1 and LF2; (c) interbedded diamicton and sand (LFA3); (d) sand lithofacies with clay/silt balls, drop stones and a pebble pavement; and (e) laminated silt and clay lithofacies (LF4) capped by diamicton (LF5). Red box indicates position of thin section taken from LFA3.





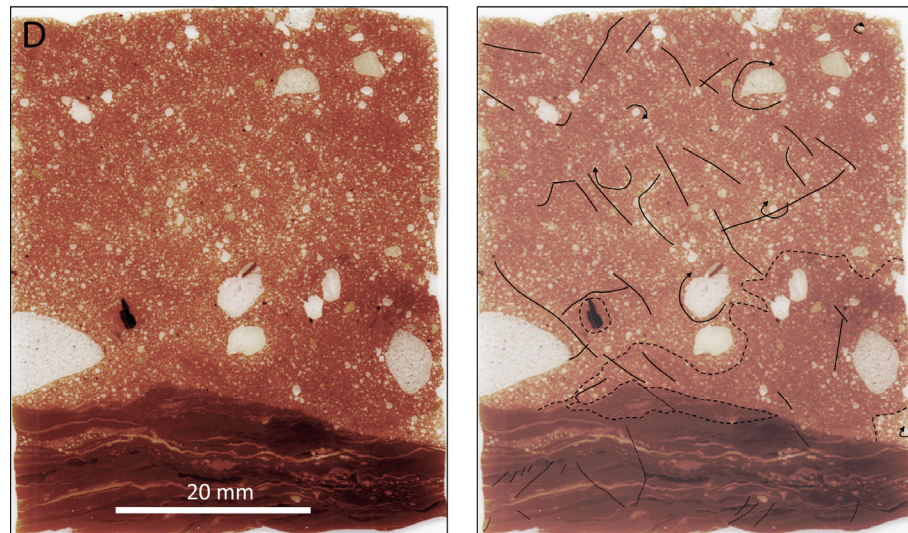


Figure 6.5: Thin sections from Blackhall Wood. All paired-images are scans of the original thin sections, while the scale bar is located at the bottom of each original image. In all images dotted lines mark the boundary between domains, while arrows indicate turbate structures, voids are delineated by shading and lines demarcate lineations/grain stacking. In image (d) in the laminated clay and silt lithofacies the lines instead indicate the position of faults: (a) LF1: lineations, grain stacking, turbate structures, adherent sediment matrixes, strong skelsepic plasmic fabric and an interconnected network of sub-horizontal and sub-vertical fissures; (b) LF2: lineations and grain stacking with some small turbate structures and a weak skelsepic plasmic fabric; (c) LFA3: interfingering sand, and diamicton of which there a number of domains. Planar and vugh voids, soft sediment pods and adherent sediment matrixes are also evident and (d) LF4 and 5: the laminated clay and silt is heavily deformed, with a series of normal faults and contortions. The boundary with LF5 is uneven, indistinct and graded with evidence of loading strictures. LF5 is characterised by two domains, with evidence of soft sediment rip up from the underlying lithofacies, soft sediment pods containing LF4, lineations, grain crushing and small turbate structures. LF5 exhibits a strong skelsepic fabric with evidence of some banding.

Overlying this lithofacies is another matrix-supported diamicton up to 2.0 m thick (LF2) characterised by a dull reddish-brown colour (5.0 YR 4/3) and massive structure (Figs. 6.3 and 6.4b). The boundary between the two lithofacies is indistinct (Fig. 6.4b) with a slight change in colour, decrease in fissility and an increase in clast content observed (Figs. 6.3 and 6.4b). A thin section taken 20 cm from the top of the LF indicates a reduction in fissures relative to LF1, although with lineations and turbate structures still present (Fig. 6.5b). Clasts in LF2 have fewer striae and are rounder in shape than LF1. The erratic content (Table 6.2) is very similar to that of LF1 with 30% greywacke, 16% Carboniferous limestone, 5% sandstone, 15% dolerite, 6% Permo-Triassic sandstone, 3% lavas from the Borrowdale Volcanic Group and 3% Southern Upland granite. However, clast macro-fabrics (Fig. 6.6) indicate a change in flow direction, from WSW-ENE in LF1 to SE-NW ( $S_1$  eigenvalues of 0.63 and 0.57) in LF2.

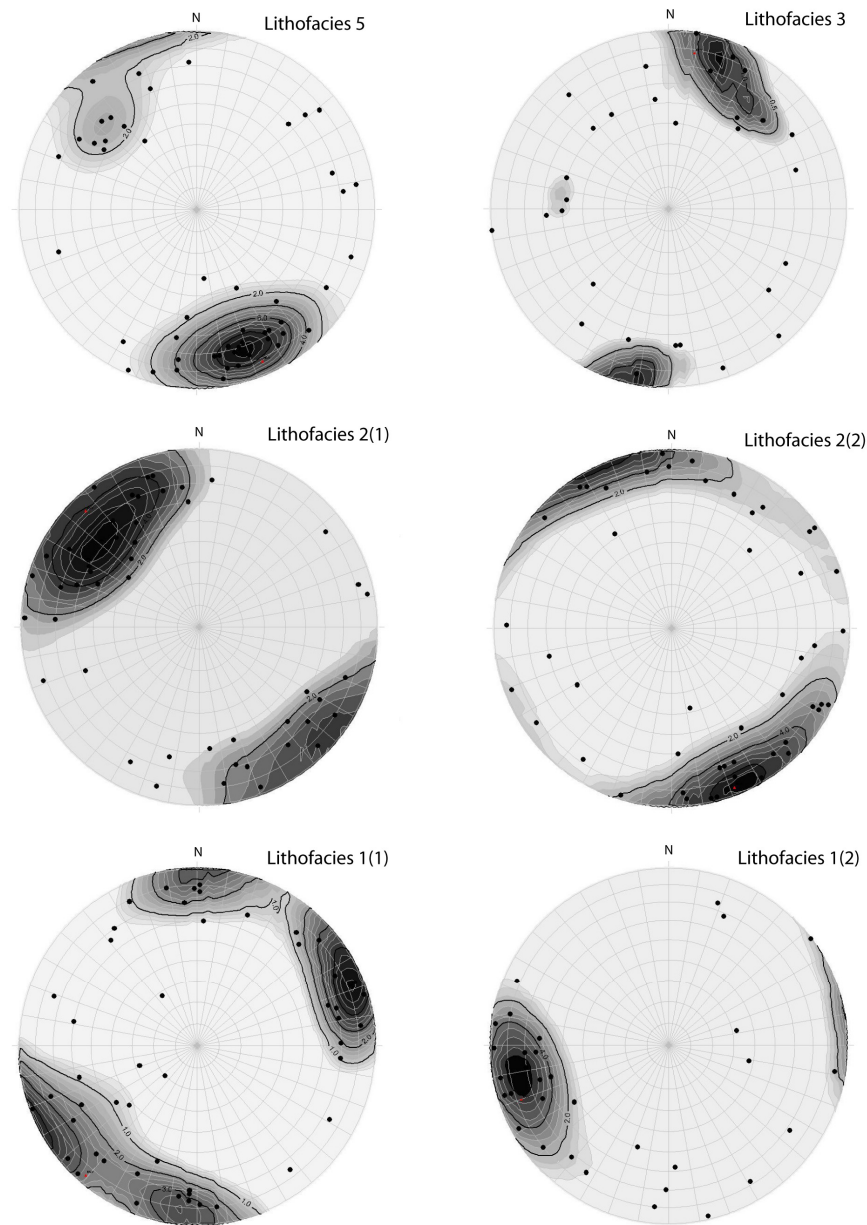


Figure 6.6: Clast fabrics for Blackhall Wood, plotted using Rockware<sup>TM</sup> with Schmidt (equal area) hemispheric projection and contoured using the Gaussian method.

#### LF1 and LF2: Interpretation

Lithofacies 1 and 2 are interpreted as subglacial traction tills (e.g. Evans *et al.*, 2006). Evidence for this includes: (a) stones that are commonly striated and have a far-travelled component to their provenance; (b) fabric orientations which have predominantly high  $S_1$  values (Benn, 1995; Evans, 2000) and girdle fabrics indicating transport in a deforming medium (Hicock & Fuller, 1995); and (c) micro-scale structures of both ductile (skelsepic fabric, turbates) and brittle (lineations, grain stacking, fissility) deformation, typical of a high-stress polyphase till (van der Meer, 1993; Menzies, 2000; Menzies *et al.*, 2006).

# CHAPTER 6: PROGLACIAL LAKE FORMATION IN THE SOLWAY LOWLANDS

The fissility exhibited by LF1, in combination with the frequency of shear planes (lineations) suggests that subglacial shear was a major process acting on the till (e.g. Hart & Boulton, 1991) possibly as a result of dewatering (Hicock & Fuller, 1995; Evans, *et al.*, 2006). The horizontal sand stringers could have formed in a low energy subglacial environment as a result of ice-bed decoupling and the development of a thin water film (cf. Piotrowski & Tulaczyk, 1999; Piotrowski *et al.*, 2001, 2002, 2004, 2006), or by glaciotectionisation of pre-existing stratified sediments (Hart & Boulton, 1991; Hart & Roberts, 1994; Hart, 1995; Benn & Evans, 1996; Evans, 2000).

	<i>Clast lithology</i>	<i>Blackhall Wood</i>			<i>Swarthy Hill</i>	<i>Maryport</i>
		<i>LF1</i>	<i>LF2</i>	<i>LF5</i>		
Metamorphic	Slate	1.19	0.28	3.53	1.64	0.79
	Schist	1.58	0.57	0.39	1.64	0.26
	Brown Quartzite	0.40	0.85	1.96	1.97	1.59
	Red Quartzite	0.40	0.28	0.00	0.33	0.00
	White Quartzite	0.00	0.57	0.00	2.62	2.65
Igneous	Carrock Fell Gabbro	0.00	0.00	0.39	0.00	0.26
	Skiddaw Granite	0.00	0.00	0.00	1.97	0.53
	Criffel Granite	0.40	0.85	1.18	0.00	0.00
	Dalbeattie Granite	1.98	1.98	0.78	1.97	1.59
	Loch Doon Granite	0.40	0.57	0.00	0.00	0.00
	Rhyolite	0.40	1.13	0.78	1.64	1.59
	Andesite	0.00	1.13	0.00	0.00	0.00
	Brown Andesite	0.00	0.00	0.00	0.00	0.53
	Borrowdale Volcanic Lava's	0.40	2.83	0.78	10.16	6.35
	Basalt	0.00	0.57	0.39	0.66	1.32
	Dolerite	8.30	15.30	7.06	4.26	6.61
	Diorite	0.00	0.00	0.00	0.00	2.12
	Felsite	2.37	1.13	1.18	1.31	1.85
	Porthyry	0.00	0.57	0.00	0.00	0.53
	Unidentified Granite	1.98	1.70	0.78	0.98	1.85
Sedimentary	Silurian Greywacke	30.83	30.31	40.00	14.75	23.54
	Permo-Triassic Sandstone	9.49	6.23	4.71	1.64	9.26
	Sandstone	12.65	5.38	14.51	32.13	13.49
	Quartzitic sandstone	1.58	2.27	1.18	2.95	6.35
	Old Red Sandstone	3.56	2.83	4.71	0.98	0.00
	Shale	0.79	0.28	0.00	0.00	0.00
	Magnesium Limestone	0.00	0.28	0.39	0.00	0.00
	Carboniferous Limestone	14.62	15.86	12.94	0.00	0.00
	Mudstone	3.16	1.98	0.39	0.00	1.85
	Siltstone	0.00	1.13	0.78	5.57	9.52
	Breccia	0.00	0.28	0.00	0.00	0.00
Mineral Veins	Quartz	1.98	0.28	1.18	0.00	1.59

Table 6.2: Average clast lithology (%) from Blackhall Wood, Maryport and Swarthy Hill, 8-64 mm. Carrock Fell Gabbro and Skiddaw Granite are sourced from the Lake District and Criffel, Dalbeattie and Loch Doon Granite are sourced from the Southern Uplands. Lava's from the Borrowdale Volcanic Group are sourced from the Lake District.



The similar provenances exhibited by the tills points to a common source region. However, the mix of different erratics from the Lake District, Southern Uplands and Vale of Eden make any interpretation of ice-flow direction based purely on provenance hard to justify. This range of lithologies from different source regions can be reconciled by the incorporation and cannibalisation of previously deposited sediments. The similar provenances exhibited between the lowermost tills and LF5 (see below) coupled with the macrofabric data suggests that the sediments were deposited by ice flowing down the Vale of Eden. The striking similarities between LF1 and LF2 suggest that they are associated with the same flow phase (convergent flow of Lake District and Southern Upland ice through the Tyne Gap) (LT1-3: Chapter 2). The macrofabric data of LF1 indicates that flow was initially north-eastwards as a tributary of the Tyne Gap Ice Stream (LT1-3: Chapter 2 and 4). Ice subsequently receded from the Tyne Gap (which ceased functioning as an ice stream) and LF2 was deposited as ice lobed-out into the Solway Basin from the Vale of Eden (thus explaining its SE-NW macrofabric). The Scottish erratics are likely to have been derived from the ‘Early Scottish advance’ of ice up the Vale of Eden (Hollingworth, 1931).

#### Lithofacies Association 3 (LFA3): Heterogeneous diamicton

##### Description

LF2 is overlain by a series of interbedded (<20 cm thick) medium-to-coarse sand and diamicton beds up to 1.1 m thick (Figs. 6.3 and 6.4c). The diamicton is similar to that of the underlying sediments (LF2) but with fewer clasts (Fig. 6.4c). The sand, which becomes increasingly dominant upwards in the LFA, is medium-to-coarse textured with some fine gravel (Fig. 6.3). Loaded contacts and convolute bedding are indicative of soft-sediment deformation (Fig. 6.4c). A clast macro-fabric taken from the diamicton displays a SW-NE orientation although with a low  $S_1$  eigenvalue of 0.49 (Fig. 6.6). A thin section taken across the boundary between the diamicton and sand lithofacies reveals a complex sequence characterised by multiple domains often organised into bands, grains with adherent sediment matrixes, soft sediment clasts, load structures, convolute bedding, both planar and vugh voids and rare examples of lineations and turbate structures (Fig. 6.5c).

The interbedded sand and diamicton grades up into massive coarse-to-medium sand and fine gravel. This lithofacies is up to 0.8 m thick and is characterised by fining upwards, silt and clay balls up to 30 cm in diameter, isolated large pebbles and a pebble layer composed of rounded-to-sub-rounded clasts (Figs. 6.3 and 6.4d). The isolated pebbles are up to 14 cm in length and tend to be more prevalent towards the top of the lithofacies, while the pebble-gravel layer, which lies 5 cm from the top of the sequence, is composed of a single layer of clasts <10 cm in length with their

long axes arranged horizontally (Fig. 6.3). LFA3 is, with the exception of the pebble-gravel layer, laterally discontinuous (Fig. 6.3).

#### LFA3: Interpretation

The interbedded diamicton and sand beds are interpreted as the product of intermittent debris flow activity. This interpretation is based on the presence of load structures, intercalated sand and diamicton, convolute bedding, multiple domains, gradational contacts and turbate structures seen both macro- and microscopically. These are indicative of deformation associated with high water contents and subject to low-intensity shear or vertical collapse; typically associated with debris flows on the foreland of glaciers (e.g. Lawson, 1979, 1981, 1982; Eyles, 1987; Lachniet *et al.*, 2001; McCarroll & Rijdsdijk, 2003; Menzies & Zaniewski, 2003). Further evidence consists of (a) bands of microscopically observed darker diamicton which form ‘tile’ structures (see Fig. 6.5c Menzies & Zaniewski (2003) consider these structures to have been formed by the pulsed movement of fine grained silts and clays via porewater, and they interpret it to be diagnostic of debris-flows (cf. Menzies & Zaniewski, 2003); (b) the geometry of the LFA, which is typical of observed debris-flow deposits (cf. Lawson, 1981, 1982); whilst (c) the upwards relationship through coarse-grained debris-flows and then glaciolacustrine sedimentation (see LF4) is indicative of a progressively ice-distal environment. The absence of high-strain shear structures such as thrust and detached folds, up-sequence increases in the strain rate or a clear principle strain axis coupled with the low  $S_1$  eigenvalues distinguish LFA3 from glacially overridden sediment (e.g. McCarroll & Rijdsdijk, 2003; Phillips *et al.*, 2008).

The debris-flow deposits grade up into medium-to-coarse sand and some fine gravel. These relatively thin deposits are massive, which would suggest an outwash event (Donnelly & Harris, 1989) close to the ice margin, or deposition by sediment gravity flows (Eyles *et al.*, 1987). Soft sediment balls composed of clay and silt within the sand and gravel are likely to be ‘rip-ups’ from shallow pools on the glacier forefield (Knight, 1999), while isolated pebbles, up to 14 cm in length, were probably cannibalised from pre-existing sub-, supra- or proglacial deposits.

#### Lithofacies 4: Laminated silts and clays

##### Description

A sharp boundary marks the upward transition from debris-flow deposits and outwash to finely laminated sediments (Fig. 6.3) comprising sub-centimetre scale dull reddish-brown (5.0 YR 5/3) clay and reddish-grey (2.5 YR 4/1) silt. Sand laminae are discernible in places, especially towards the top of the unit. The upper contact of these laminated sediments with the overlying diamicton (see below) is undulatory, and as a result, the thickness of the laminated unit ranges from a few

centimetres up to 1.3 m. Small drop-stones of fine gravel are infrequently found throughout the laminated sequence.

Micromorphological description of the laminated sediments follows the procedure outlined by Palmer *et al.* (2008a, b), where it is possible to resolve different lamination sets, which are defined by the presence of alternations between silt/sand (coarse component) and clay (fine component) laminations. These elements of a coarse component, sometimes containing multiple laminations of silt and sand, and a fine component of clay are identified as a couplet (Palmer *et al.*, b). Three microfacies groups are identified within the laminated section. Evidence of deformation is also described.

Microfacies 1 (0 – 7.52 cm & 21.89 – 35.18 cm from base of lamination unit) comprises 66 couplets. These consist of alternations between a fine component of a single lamination of predominantly clay and a coarse component of predominantly silt or very fine sand laminations (Fig. 6.7a, b). The clay has sharp upper and lower contacts, is horizontally aligned and grades from very fine silt to clay. Under cross-polarised light, the clay shows high birefringence fabric. The coarse component has multiple, normally-graded silt laminations with sharp, upper and lower contacts (Fig. 6.7a). A maximum of seven graded laminations are identified in the coarse component of the couplet, in this microfacies group. This is sometimes overlain by a single massive, sand or silt lamination, which has a sharp contact to the overlying fine component (Fig. 6.7b). The coarse component is thicker than the fine component. Couplets of the fine and coarse component have an average thickness of 3.1 mm, although there is a wide range (0.30 – 9.74 mm), and considerable variation (standard deviation 2.3 mm) throughout the group.

Microfacies 2 (7.52 – 21.89 cm) comprises 76 couplets and is similar to microfacies 1, but with three key differences. The first is the coarse component is generally composed of a single layer of massive, medium silt, with occasional multiple, very fine, lamination present (Fig. 6.7c). The second is that the thickness of the coarse component is equal to the fine component. As a result, the third difference is that this microfacies group has thinner couplets than group 1 (1.72 mm on average) with only three sets exceeding 3.5 mm.

Microfacies 3 (35.18 – 101.94 cm) comprises 119 couplets of silt and clay, similar to microfacies 1 except that the laminations are much thicker (5.61 mm on average) and the silt and clay layers are of equal thickness (Fig. 6.7d). Up to 15 laminations have been measured within the coarse component and large amounts of coarse material have also been observed (Fig. 6.7e).

Evidence of deformation is apparent towards the top of the lithofacies (101.94 – 127.41 cm) and also through microfacies 2 (11.40 – 12.69 cm). The first area is in microfacies group 2, which is composed of wavy, occasionally truncated, thin clay and silt laminae, with some sub-vertical birefringent fabric development. The second zone of deformed laminations is toward the top of the

sequence and is characterised by normal faults, soft-sediment intra-clasts, fabric development within silt bands, sheared boundaries and folding around drop-grains. Further up the sequence, the laminations become heavily contorted and folded (Fig. 7f), with more coarse material and vertical injections of sediment. The very top of the sequence is relatively undisturbed and is composed of thin horizontally laminated clay and silt.

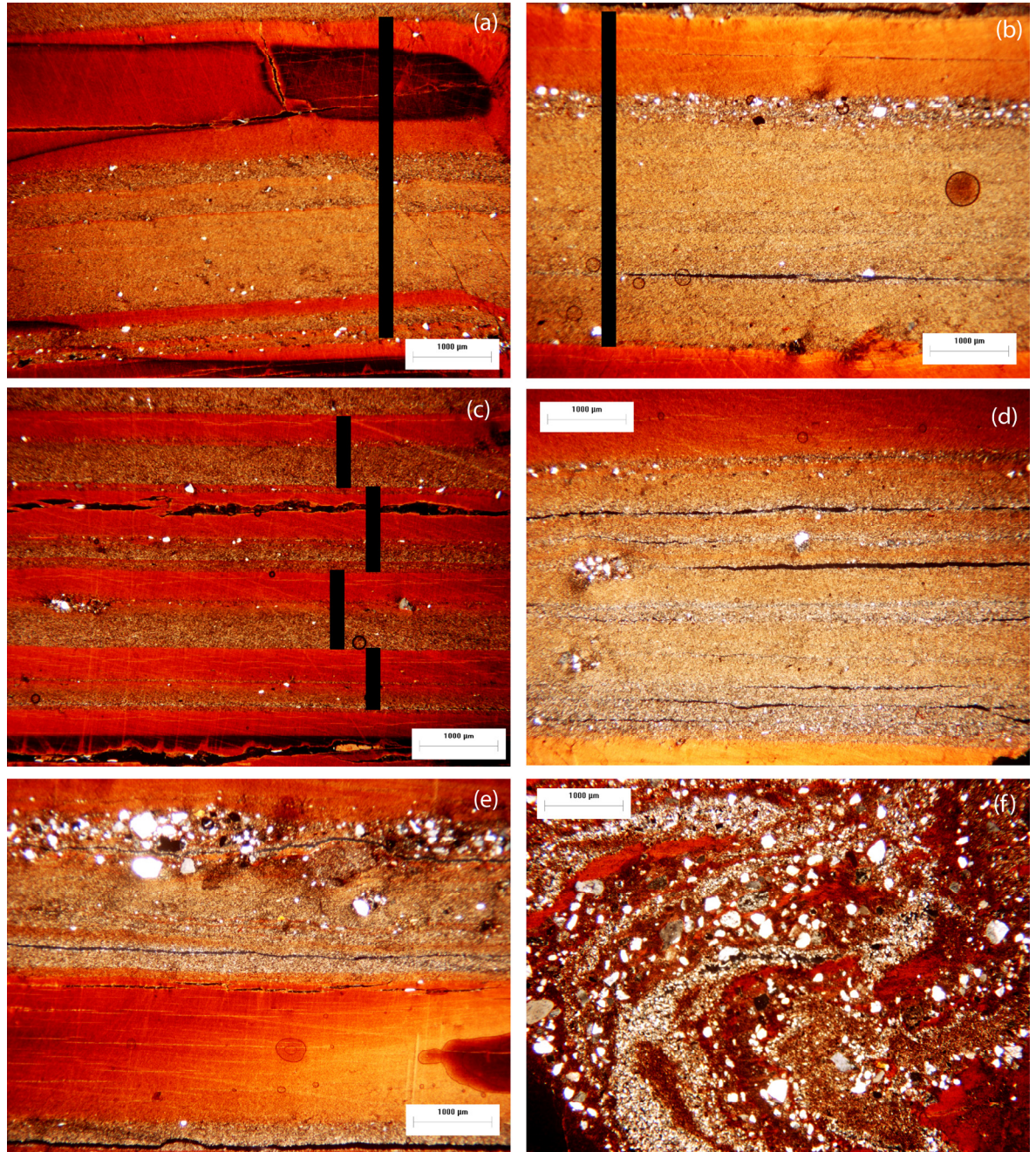


Figure 6.7: Microphotographs of typical laminated sediment from Blackhall Wood. All images have a scale bar in one of their corners and were taken under cross-polarized light. Black bars delimit one varve. (a) microfacies group 1: note the sharp contacts between the top of the fine lamina and beginning of the coarse component, which is characterised by multiple, normally graded silt laminations; (b) microfacies 1: note the coarser laminae of very fine sand and silt towards the top of the coarse component, which is normally graded; (c) microfacies 2: very thin regular laminae with sharp boundaries; (d) microfacies 3: thick laminae which are



< Figure 6.7 (continued)... characterised by multiple normally graded laminations within the coarse component; (e) microfacies 3: characterised by coarser very fine sand and silt towards the top of the coarse component; and (f) deformed laminae: heavily contorted and folded laminae, with large amounts of coarse material.

#### LF4: Interpretation

Microfacies 1-3 exhibit a number of distinctive properties which allow them to be interpreted as clastic varves. These include alternations between a coarse-grained component (silt and very fine sand), which has multiple graded laminae, and a fine-grained component (clay lamina). There is a sharp contact between both the coarse component and the fine component, and between the fine and coarse component. The coarse component records multiple input events as underflows to the lake basin during the summer period, with a sharp contact suggesting a break in sedimentation, possibly related to the autumn overturn (Delaney, 2007). The fine component of the clay laminae which succeeds the coarse component forms during the winter through the settling of suspended material in the water column, when the lake water freezes, which inhibits the generation of wind-driven currents in the lake basin that normally causes the re-suspension of clay during the summer (Smith, 1978; Smith & Ashley, 1985; Ringberg & Erlstrom, 1999; Delaney, 2008; Palmer, 2008b). The sharp contact between the fine winter lamina and the succeeding coarse summer laminae indicates the renewed transport of sediment to the lake basin during the spring/summer melt season (Ashley, 1975). The varve thickness characteristics throughout the sequence are indicative of sedimentation in distal zones of the lake basin (Ashley, 1985; Smith & Ashley, 1985; Ringberg & Erlstrom, 1999). The lack of coarse-grained dropstones could also indicate ice-distal conditions, or alternatively shallow water. There are subtle differences in the microstructures observed within the summer layers which are highlighted below.

Microfacies 1 is characterised by micro-laminated silt layers, with up to 7 laminations of graded sediment. These micro-laminae relate to episodic density current underflows which could result from the release of meltwater from different sources (Palmer *et al.*, 2008b), increases in meltwater discharge (Ringberg & Erlström, 1999), surge currents from slumping sediment (Smith & Ashley, 1985) or rainfall events (cf. Delaney, 2007). The coarse material deposited at the top of the silt component are potentially attributed to wind generated bottom currents produced during storms in the late melt season (Ringberg & Erlström, 1999), precipitation events during the Autumn (Smith, 1978) or an increase in aeolian deposition (Lamoureux & Gilbert, 2004). The sharp contact at the base of the silt layer marks the sudden onset of underflow sedimentation during the spring melt season, while graded structures in the summer layer (coarse component) indicate complete sedimentation from each discrete sediment pulse that enters the lake basin.

Microfacies 2 has fewer laminations of sediment within the summer layer suggesting that only extreme underflow events entering the lake have sufficient velocity to reach this point in the basin, and therefore, the varve thickness is decreased (Ringberg & Erlström, 1999). This could be the product of reduced glacier extent within the catchment (Leonard, 1986) and the similar thickness of the coarse and fine component may relate to increased dispersion of sediment from the input region or a reduced melt season.

Microfacies 3 is interpreted in a similar way to microfacies 1. However, a greater average thickness of the couplets, number of laminae within the coarse component and presence of coarse grained material suggests that the microfacies was relatively more ice-proximal, although still in a distal part of the basin.

Deformation exhibited within the sequence comprises faults, folds, water escape structures, loading, soft-sediment intra-clasts and sheared boundaries thus indicating that the varves were subject to post-depositional deformation. The prevalence of these structures at the top of the sequence coupled with the upwards increase in strain signature indicates that the deformation may be related to the emplacement of the capping diamicton (see below).

A total number of 261 years are recorded in the sequence and the thickness varies through time (Fig. 6.8). Initially the thickness decreases, marked by a transition from microfacies 1 (24 years) to microfacies 2. Deposition of microfacies 2 occurred over a period of 76 years before varve thickness increases through microfacies 1 (41 years) and then the microfacies 3 (119 years) at the top of the sequence, where the thickest varves are observed (Fig. 6.8). These changes in varve thickness may reflect variations in the position of the ice margin such that ice retreats allow the formation of glaciolacustrine varve sediments. Subsequently ice extent increases within the catchment until eventually the lake sediments are overridden indicated by the deformation of the sediments at the top of the sequence (see below). The varves subjected to deformation were not included within the time sequence due to the implicit difficulties of producing an accurate count when the laminae have been subject to shortening, compression and possible repetition due to folding and shearing. This implies that the varve count probably under-estimates the duration of the lakes existence.



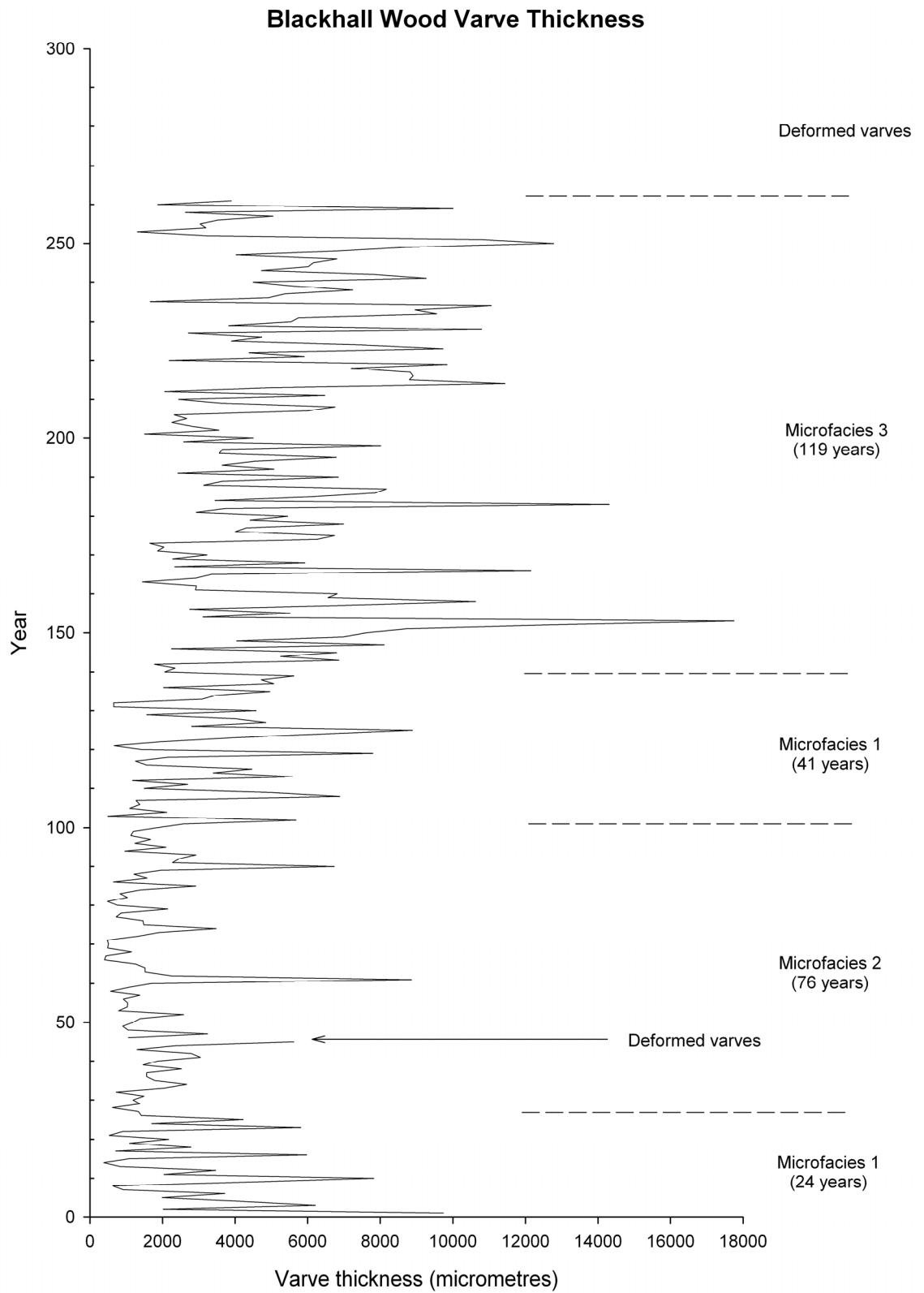


Figure 6.8: Varve thickness and count.

Lithofacies 5: Uppermost diamictonDescription

The uppermost lithofacies consists of a massive, dull reddish-brown (5 YR 4/4) matrix-supported diamicton (Fig. 6.4e) up to 20 m thick that forms the bulk of the drumlin (Fig. 6.3). The diamicton is clast-rich with sub-rounded to rounded morphologies and rare striae. Its erratic content (Table 6.2) is similar to LF1 and LF2 with 40% greywacke, 15% sandstone, 13% Carboniferous limestone, 7% dolerite, 5% Permo-Triassic sandstone, 4% slate, 2% Southern Upland granites and 1% lavas from the Borrowdale Volcanic Group. The boundary between the diamicton lithofacies and the underlying laminated clay and silt is erosional and undulatory (Fig. 6.3). A thin section taken across this boundary displays deformed clay and silt, with normal faulting and folding (Fig. 6.5d). Rip-up clasts of clay have been incorporated into the diamicton unit and a darker diamicton domain towards the boundary edge shows evidence of strong banding and skelsepic plasmic fabrics (Fig. 6.5d). The diamicton within the thin section is dominated by dipping lineations, crushed grains, pressure shadows and occasional turbate structures. A clast fabric taken near the base of the lithofacies (Fig. 6.6) indicates a SE-NW orientation with an  $S_1$  eigenvalue of 0.62.

LF5: Interpretation

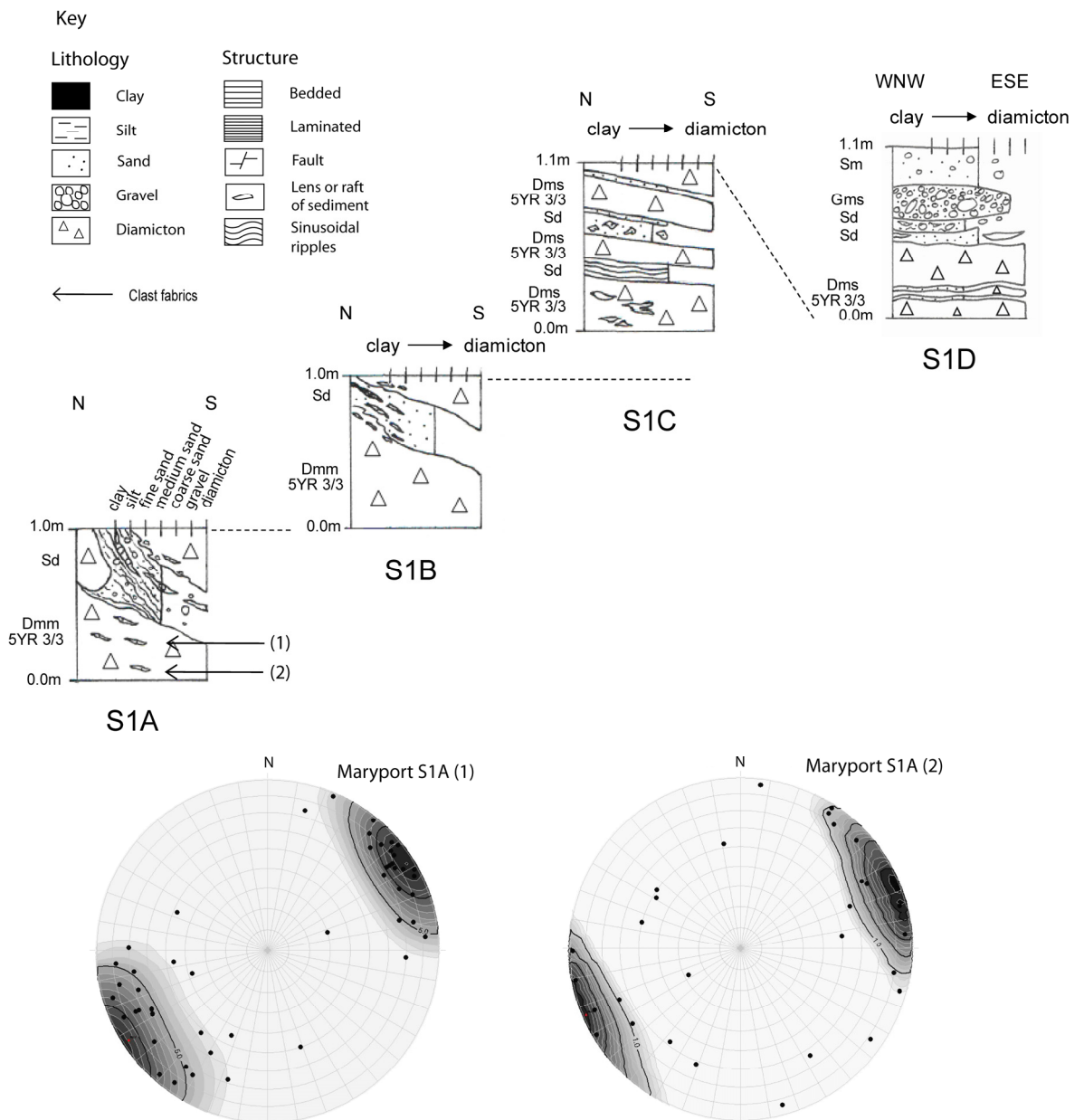
LF5 is interpreted as a subglacial traction till (*sensu* Evans *et al.*, 2006) based on the following evidence: (a) the large thickness (20 m) and massive nature of the diamicton which has been moulded into a SE-NW orientated drumlin, indicating glacial overriding (Menzies, 1979; Boulton, 1987; Hart, 1997) and probably some localised folding and stacking (Evans *et al.*, 2006); (b) the fabric orientation is consistent with the drumlin's form, while the  $S_1$  eigenvalue indicates strong streamlining of clasts in the direction of ice motion (Benn, 1995; Evans, 2000); (c) clasts contain a far-travelled provenance; (d) the diamicton has a sharp, erosional boundary with the underlying, partially deformed varved clay and silt lithofacies (cf. Evans *et al.*, 2006); (e) clay and soft sediment clasts have been incorporated into the till matrix, indicating cannibalisation of the underlying varved sediments and incorporation into the diamict (e.g. Hicock & Dreimanis, 1989; Benn & Evans, 1996; Hooyer & Iverson, 2000); and (f) the thin section contains lineations, turbate structures, skelsepic and banded plasmic fabrics and pressure shadows typical of a polyphase till that underwent brittle and ductile deformation (van der Meer, 1993; Menzies, 2000; Hiemstra & Rijdsdijk, 2003; Menzies *et al.*, 2006). Indeed the propensity of lineations and low porosity relative to other microstructures suggests that brittle shear was the dominant process near the base of the diamicton (Menzies, 2000; Menzies *et al.*, 2006).

The sub-rounded to rounded clast morphologies coupled with the diverse clast provenance and rare striae suggests that the till has incorporated glaciofluvial/glaciolacustrine sediment and diamicton of previously deposited sediment (Hicock & Dreimanis, 1989; Hicock & Fuller, 1995; Evans,

2000). The orientation of the drumlin indicates that ice flow was northerly moving down the Vale of Eden before swinging west into the Irish Sea Ice Basin (LT5: Chapter 2; Livingstone *et al.*, 2008).

#### 6.4.2 Maryport:

Maryport (NY 038 378) is located on the Cumbrian coast (Fig. 6.1), with exposures cut into the southwest flank of a large, NE-SW orientated drumlin (56 m O. D.). The drumlin forms part of a flowset which sweeps round the northern edge of the Lake District (LT5: Chapter 2; Fig. 6.2). The underlying bedrock is Permo-Triassic sandstone. The sedimentary lithofacies are depicted diagrammatically in Fig. 6.9 and described below.



< Figure 6.9: Maryport section logs and fabrics.

#### Description:

The exposure displays one lithofacies, which is up to ~ 20 m thick and comprises a matrix-supported, clast-rich reddish (5 YR 3/3) diamicton (Figs. 6.9 and 6.10a). Clasts have sub-angular to sub-rounded morphologies and rare striae. The principle lithologies (Table 6.2) are Silurian greywacke (24%), sandstone (13%), siltstone (10%), Permo-Triassic sandstone (9%), dolerite (7%), lavas from the Borrowdale Volcanic Group (6%), and quartzitic sandstone (6%), with 2% Southern Upland granite and 1% Lake District granite also observed. Two clast fabrics taken in the upper exposures (Fig. 6.9) indicate a NE-SW orientation with  $S_1$  eigenvalues of 0.77 and 0.60. A thin section from the diamicton exhibits lineations, turbate structures, skelsepic plasmic fabrics and adherent sediment matrixes.

The diamicton contains a series of interbedded and folded sand and fine-gravel beds (Figs. 6.9 and 6.10a, b), sand stringers and pebble gravel clusters (Fig. 6.10c). Two large sand open folds up to 1.0 m thick exposed about half way up the drumlin (Figs. 6.9 and 6.10a) show evidence of primary stratification and are attenuated-out towards the S/SW, with the hinge-line orientated towards the NE. This stratification has been partially deformed, with centimetre-scale folds, faults and clay stringers. Boundaries with the diamicton are sheared and mixed. In the uppermost exposures the diamicton is inter-bedded with slightly folded/faulted sand stringers, sand lenses, fine-pebble gravel beds (up to 50 cm) and partially stratified sand beds (up to 50 cm) (Figs. 6.9 and 6.10b). These upper diamictons are very clast rich with sub-rounded to rounded morphologies. A vertically orientated, clast-supported pillar of pebble-gravel 10 cm wide by 40 cm high was also observed (Fig. 6.10c)

#### Interpretation:

The lithofacies at Maryport are consistent with an origin as a glacioteconite (*sensu* Evans *et al.*, 2006) derived from the glacioteconisation of underlying glaciofluvial/glaciolacustrine sediments (e.g. Broster, 1991; Hart & Roberts, 1994; Benn & Evans, 1996; Evans, 2000). The surface of the diamicton has been moulded into a drumlin, with the high  $S_1$  eigenvalues of the diamicton suggesting subglacial streamlining of a deformable substrate (Benn, 1995; Evans, 2000; Evans *et al.*, 2006). Sand and gravel beds, which have been preserved as competent masses, have been rafted upwards into the diamicton and folded and attenuated in the direction of ice flow (cf. Hart & Boulton, 1991; Hart & Roberts, 1994) whilst rounded to sub-rounded clasts and adherent sediment matrixes suggest the cannibalisation of underlying sediment to form at least part of the till matrix (e.g. Hicock & Dreimanis, 1989; Hicock & Fuller, 1995; Evans, 2000). Thus, the clast provenance is likely to have been partially derived from the underlying glaciofluvial/glaciolacustrine sediments. The sand lithofacies show some preservation of primary stratification. The presence of



thinly bedded (<50 cm), massive, clast-rich diamicton, with sand lenses and thin beds of fine-to-coarse sand, in association with the gravel units, supports a debris-flow interpretation (e.g. Eyles, 1987). The vertical pillar of clast-supported, clast rich pebble-to-cobble gravel is likely to be a clastic dyke derived from a gravel-rich unit (cf. van der Meer *et al.*, 2009). This, coupled with the streamlining of the diamicton and the sand, and the attenuated sand structures all indicate that there were high pore water pressures (e.g. Broster, 1991) and that simple shear was a key mechanism in the creation of internal structure (e.g. McCarroll & Rijdsdijk, 2003).



< Figure 6.10: Photographs from Maryport and Swarthy Hill: (a) Maryport: diamicton and sand folds; (b) Maryport: folded and interbedded sand; (c) Maryport: pebble-cobble gravel cluster; (d) Swarthy Hill: fine sand beds with clastic dykes leading into gravel beds; (e) Swarthy Hill: diamicton lithofacies interbedded with sand; and (f) Swarthy Hill: clastic dykes.

#### 6.4.3 *Swarthy Hill:*

Exposures at Swarthy Hill (NY 069 403) are located in the southwest flank of a NE-SW orientated drumlin (31 m O. D.). The drumlin is part of an arcuate flowset which sweeps round the northern edge of the Lake District (LT5: Chapter 2; Figs. 6.1 and 6.2). The underlying bedrock geology is composed of Permo-Triassic sandstone. The sedimentary lithofacies are depicted diagrammatically in Fig. 6.11 and described below.

##### Description:

The exposures at Swarthy Hill are dominated by fine sand beds up to 1.5 m thick (Figs. 6.10d and 6.11). These are chiefly massive in structure with some fine-to-cobble gravel (maximum diameter 20 cm), occasional silt and clay, and ripple structures (orientated towards the SE/ESE). Where present, pebble-to-cobble gravel is dispersed throughout the sand. The sand has been heavily deformed (Fig. 6.10d), with clay and silt stringers, clay pods, diamicton inclusions, fold structures, reverse and normal faults (with centimetre-scale offsets), flame structures, convolute bedding, coarse sand injections and cross-cutting laminae composed of fine sand, clay and silt.

Overlying the sand lithofacies are beds of up to 1.0 m thick matrix-supported, massive fine-to-cobble gravel, comprising sub-rounded to rounded clasts up to 30 cm diameter (Fig. 6.11). Thin (<5 cm) diamicton and fine sand units are inter-bedded with the gravel (Fig. 6.11). The gravel lithofacies has a loaded lower boundary.

Interbedded within the deformed sand is a thin (up to 0.5 m thick) bed of matrix supported, dull reddish-brown (5 YR 4/4) diamicton (Fig. 6.10e) that dips towards the southeast. The diamicton has a sandy composition with rounded to sub-rounded clasts up to 30 cm in diameter, indistinct sheared boundaries and a NW-SE fabric ( $S_1$  eigenvalue of 0.48) (Fig. 6.11). The provenance of the diamicton is very similar both to the subglacial till identified at Maryport and the sand and gravel exposed at Swarthy Hill (Table 6.2). This includes 32% sandstone, 15% Silurian greywacke, 10% lavas from the Borrowdale Volcanic Group, 6% siltstone 4% dolerite, 2% Scottish granite and 2% Lake District granite. Thin units of diamicton (<10 cm thick) are also interbedded within the gravel lithofacies, and as stringers and pods within the sand (Fig. 6.11).

Sheets of sub-vertically and sub-horizontally aligned fine-grained sand, with sand, silt and clay laminae up to 5 cm wide are observed throughout the exposure (Figs. 6.10f and 6.11). These



structures have consistent widths, although with tapered-out, dendritic ends (Fig. 6.11) and occasional ‘burst-out’ (Rijsdijk *et al.*, 1999, van der Meer *et al.*, 2009) structures (Fig. 6.10f). The laminae run parallel to the walls of these structures and often cross-cut each other. These structures divert around clasts, finger-up into diamicton lithofacies and run vertically into the gravel beds.

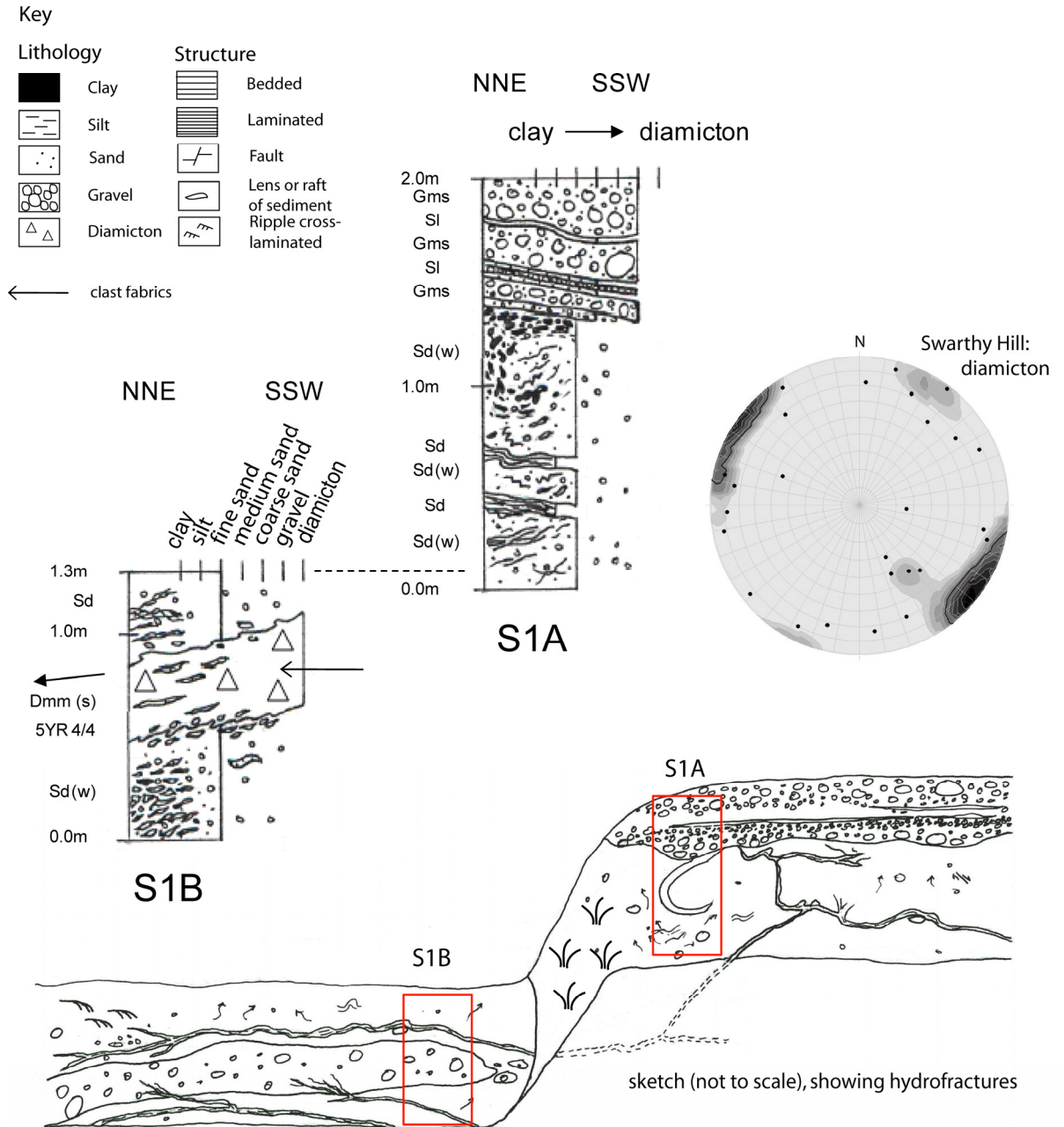


Figure 6.11: Swarthy Hill section logs, sketch of clastic dykes and fabrics.

Interpretation:

The stratified sand and gravel lithofacies are thought to have been initially deposited as proglacial glaciofluvial and glaciolacustrine sediments in an ice proximal location, with dropstones indicating ice-rafting into a standing water body. Infrequent ripples suggest a SE/ESE flowing palaeocurrent, while the sheared interbedded diamicton is probably a debris flow (Lawson, 1981, 1982). Partial destruction of the primary structure by extensive deformation and dewatering clearly demonstrates subsequent overridding and glaciotectonisation of the pre-existing sediments into a drumlinoid form (e.g. Krüger & Thomsen, 1984; Boulton, 1987; Boyce & Eyles, 1991; Menzies & Brand, 2007). Although the sediments are composed of stratified sand, gravel and diamicton, a potential genesis as a lee side cavity fill is ruled out due to: (a) the lack of distinctive dipping stratification (cf. Hillefors, 1973; Dardis *et al* 1984; Dardis, 1985; Levson & Rutter, 1989a,b); (b) the deposits show extensive evidence of deformation which is absent in the cores of cavity-fills (Dardis *et al* 1984; Dardis, 1985); (c) diamicton debris flows present within the sequence are interpreted as subaerial rather than the subaqueous flows diagnostic of cavity-fills, due to the lack of internal stratification/lamination (Evenson *et al.*, 1977), the thin geometry, sheared boundaries and positioning between deformed sand lithofacies (Lawson, 1981, 1982); and (d) the ripple structures are orientated at right angles to the orientation of the drumlin, which is at odds with the cavity-fill model of water flow into the cavity, and indeed suggests a mode of deposition independent of drumlin formation.

During glaciotectonisation the sediment is inferred to have been subject to fluctuating pore water pressures. This is consistent with: (a) centimetre-scale coarse-sediment injections (Fig. 6.9d), interpreted as water escape structures caused by pervasive de-watering; and (b) the bundles of sub-horizontally and sub-vertically orientated fine-grained laminae, with occasional burst-out structures, which are interpreted as clastic dykes (e.g. Rijdsdijk *et al.*, 1999; van der Meer *et al.* 1999, 2009; Kjaer *et al.*, 2006). Clastic dykes are thought to form by hydro-fracturing as a result of changing water pressures (van der Meer *et al.*, 2009). The clastic dykes cut across multiple lithofacies thus indicating that dewatering must have occurred post-depositionally.

#### 6.4.4 Borehole stratigraphy:

Borehole data gathered from the M6 motorway between Penrith and Carlisle (Fig. 6.1) augments observations from this study and can be compared and correlated with previous research (Goodchild, 1875; Dixon *et al.*, 1926; Huddart, 1970; Arthurton & Wadge, 1981) in order to better define the regional stratigraphy. Figure 6.12 presents the locations and thicknesses of the laminated sediments and Figure 6.13 displays the general stratigraphy of the Vale of Eden. The thickness of the upper diamicton is also recorded (Fig. 6.12c). Evidence for the tripartite sequence is

commonplace in the lower reaches of the Vale of Eden (especially in the Petteril and Caldew Valleys) below ~ 70 m (Figs. 6.12 and 6.13). Two firm to stiff reddish-brown diamictons have been recorded separated by up to 15 m of sand, gravel and laminated clay/silt lithofacies (Fig. 6.13). Above 70 m O. D., interbedded sand with some laminated clay/silt is still evident at Calthwaite, Langwathby and Low Dyke (Figs. 6.12b and 6.13). Some evidence of a tripartite sequence is also observed to the west of Blackhall Wood, with laminated clays observed in boreholes at Thursby (Fig. 6.12) and middle sands exposed along the River Wampool (Dixon *et al.*, 1926). However, borehole evidence seems to suggest that most drumlins comprising the westerly orientated flowset (LT5: Chapter 2) tend to be diamicton cored (with a capping unit of sand and gravel). The heights of the laminated clay-silt lithofacies within the stratigraphic sections generally range between 46-54 m O. D, although they are also infrequently recorded higher up the Vale of Eden (e.g. 133 m O. D. at Low Dyke) (Figs. 6.12b and 6.13). Thicknesses of the laminated clay and silt range widely from <0.5 to 5.2 m, with the thickest exposures exposed in the Petteril Valley below 100 m (Fig. 6.12b). There seems to be a general thinning west and east of the Petteril Valley, although high values are persistent northwards towards Carlisle (Fig. 6.12b). The thickness of the upper diamicton is variable, ranging from less than 5 m to over 20 m (Fig. 6.12c). The general up-down trend in thickness is probably a function of drumlins vs intervening hollows (Fig. 6.12c), although it is interesting to note a couple of particularly thick sequences of the upper diamicton at Calthwaite and Foul Bridge (Figs. 6.12c and 6.13). The large thickness of diamicton at Foulbridge is of particular importance as it corresponds with a faint ridge which can be traced to the west and east along the edge of a lobate flowset extending northwards into the fringe of the Solway Lowlands (Fig. 6.2).

Stratigraphic evidence derived from prior investigations and borehole logs can be correlated to the field section at Blackhall Wood. The uppermost dull reddish-brown till at Blackhall Wood is ubiquitous, being observed throughout the study area (Dixon *et al.*, 1926; Trotter, 1929; Hollingworth, 1931), and making up the drumlinoid features that cover the Vale of Eden. The middle sand, silt and clay lithofacies, which are often found sandwiched between diamicton units, exhibit similar heights and characteristics, with a propensity for laminated sequences in close proximity to Blackhall Wood. Thus, the middle sand, gravel and laminated clay-silt lithofacies can be correlated with some confidence. The laterally discontinuous, lower-reddish brown diamicton probably correlates to the lowermost Blackhall Wood tills (LF1 and LF2). Dixon *et al.* (1926) highlighted the lithological similarities between the lower and upper tills, which both have a reddish-brown colour. In the past, the lower diamicton has been correlated with an early (Late Devensian) Scottish advance across Stainmore (e.g. Dixon *et al.*, 1926; Hollingworth, 1931). However, evidence produced in this study would suggest that, whilst the erratic support for a Scottish-sourced ice flow is compelling, both the lower (LF1 & LF2) and upper tills (LF5), in the Solway Lowlands, were deposited from ice moving down the Vale of Eden.

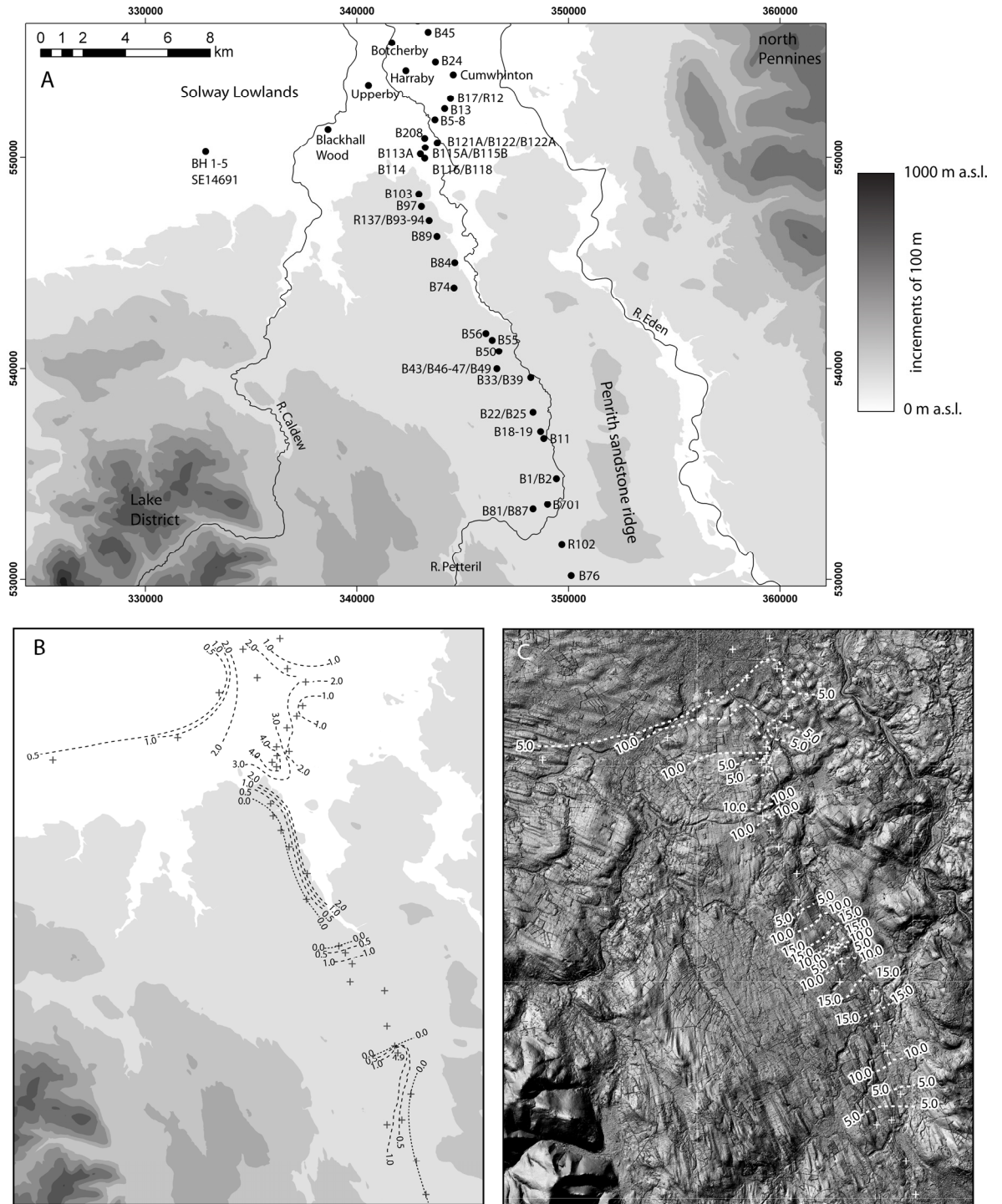


Figure 6.12: (a) Field sites, borehole locations and stratigraphic sections from prior research (Goodchild, 1875; Dixon *et al.*, 1926; Huddart, 1970; Arthurton & Wadge, 1981); (b) contour map of the thickness of laminated clay and silt within the Vale of Eden (data from Figs. 6.11 and 6.13), mapped against topography; (c) contour map of the thickness of the upper diamicton (data from Figs. 6.11 and 6.13) mapped against the NEXTMap DEM.

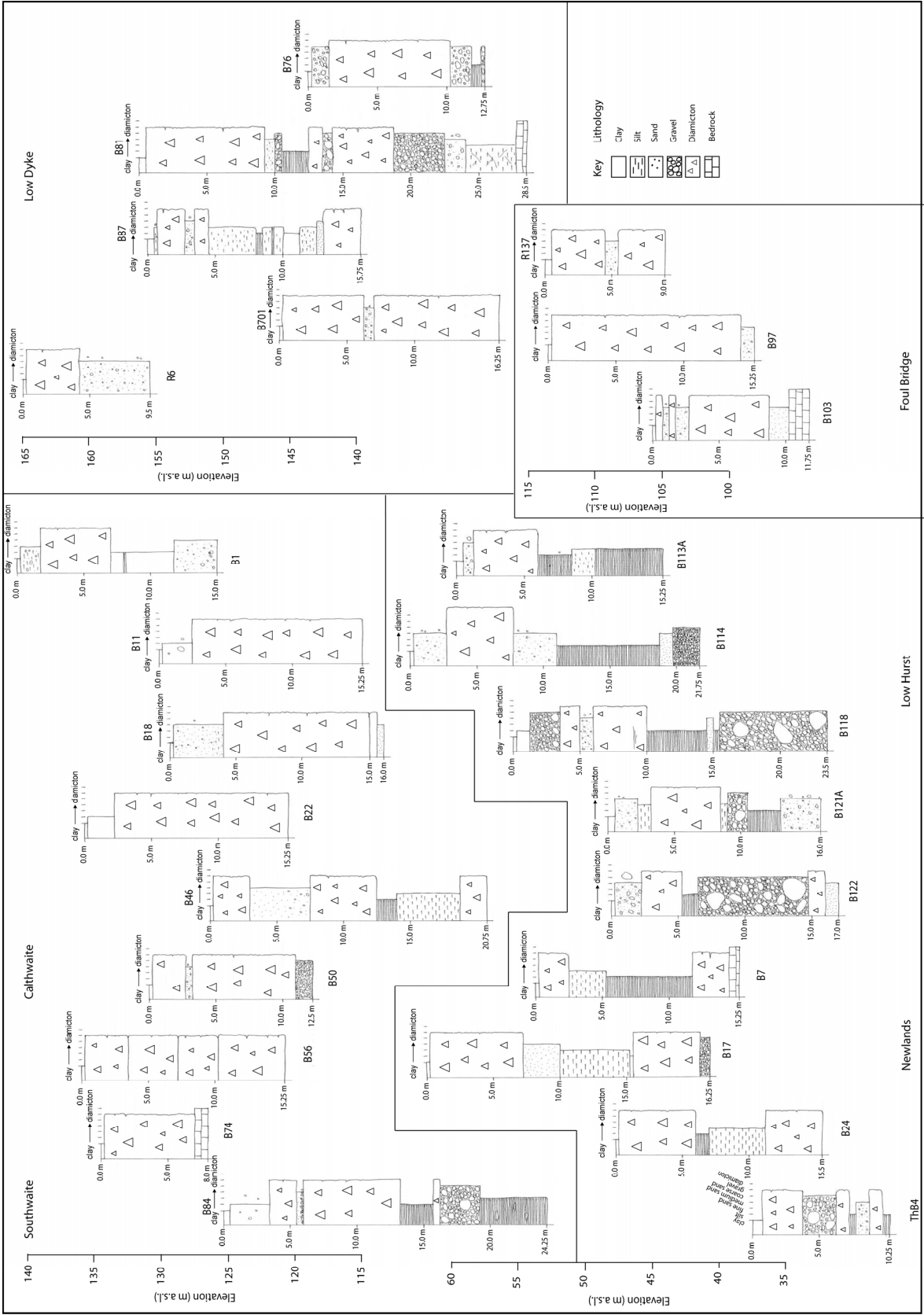


Figure 6.13: Stratigraphic logs from key sites through the Vale of Eden. See Fig. 6.12 for detailed location map of boreholes.

## 6.5 Discussion

The sedimentary and stratigraphic data obtained during this study indicates that initial deglaciation of the Solway Lowlands resulted in an ice-free enclave with ice situated in the Solway Firth promoting the rapid development of a glacial-lake, which we refer to as ‘Glacial-Lake Blackhall Wood’. Varve records suggest that this lake persisted for at least 261 years; however, the truncated upper boundary and deformed varves mean that it probably existed for longer. Re-advance of ice over the glaciofluvial and glaciolacustrine sediments produced the characteristic tripartite sequence, with the Irish Sea Ice Stream drawing Scottish and Lake District ice westwards into the Irish Sea Basin and leaving behind what is interpreted to be a fast-flow signature constrained by hummocky terrain and ribbed moraine along the northern margin of the ‘flowset’ (Figs. 6.2 and 6.14) (cf. Ross *et al.*, 2009). A subglacial interpretation for the tripartite sequence (cf. Huddart, 1970; Goodchild, 1875, 1887) is rejected here based on the presence of clastic varves and their association with both debris-flow and outwash deposits at the Blackhall Wood site. Incorporation of Scottish erratics within the lowermost till units at Blackhall Wood (LF1 and LF2) supports previous research that proposed an initial southwards flow of Scottish ice up the Vale of Eden and across the Stainmore Gap (e.g. Goodchild, 1875; Hollingworth, 1931). However, the fabric and provenance data can be reconciled with geomorphic evidence (cf. Livingstone *et al.*, 2008) that suggests that these tills were deposited during ice-flow convergence on the Solway Lowlands from the Southern Uplands and the Lake District, before moving eastwards through the Tyne Gap (Fig. 6.14a). As the ice subsequently retreated out of the Tyne Gap it lobed-out into the Solway Lowlands (LF2: Blackhall Wood). These tills are therefore thought to relate to flow phases LT1-3 (Livingstone *et al.*, 2008; Chapter 2, Fig. 2.5) with Table 6.3 illustrating where they fit within the regional Event Stratigraphy.

The debris-flow deposits at Blackhall Wood record an increasingly ice-distal environment as the Solway Lowlands became deglaciated (Table 6.3). Fieldsites at Swarthy Hill and Maryport, which show evidence of glaciotectionised proglacial glaciofluvial and glaciolacustrine sediment, indicate that ice retreated west of the Cumbrian coast. Therefore, a stretch of at least 42 km along the Solway Lowlands became deglaciated either concurrently or time-transgressively, in relation to changes in the ice-sheet’s configuration (Fig. 6.14b). Boreholes along the M6 show evidence for extensive deglaciation within the Vale of Eden (Fig. 6.14b). However, the debris-flows composed of diamicton and outwash at Blackhall Wood indicate that ice was in close proximity. Indeed the Solway Lowlands probably formed an ice-free enclave, constrained to the west by the still active Irish Sea Ice Stream (e.g. Merritt & Auton, 2000). Tripartite sequences north of Carlisle and along



the Scottish border have been correlated with the latter Scottish re-advance (flow phase SF1: Chapter 2) (cf. Stone *et al.*, in press) with the corollary being that ice probably maintained a permanent presence within this region during the earlier retreat (Fig. 6.14).

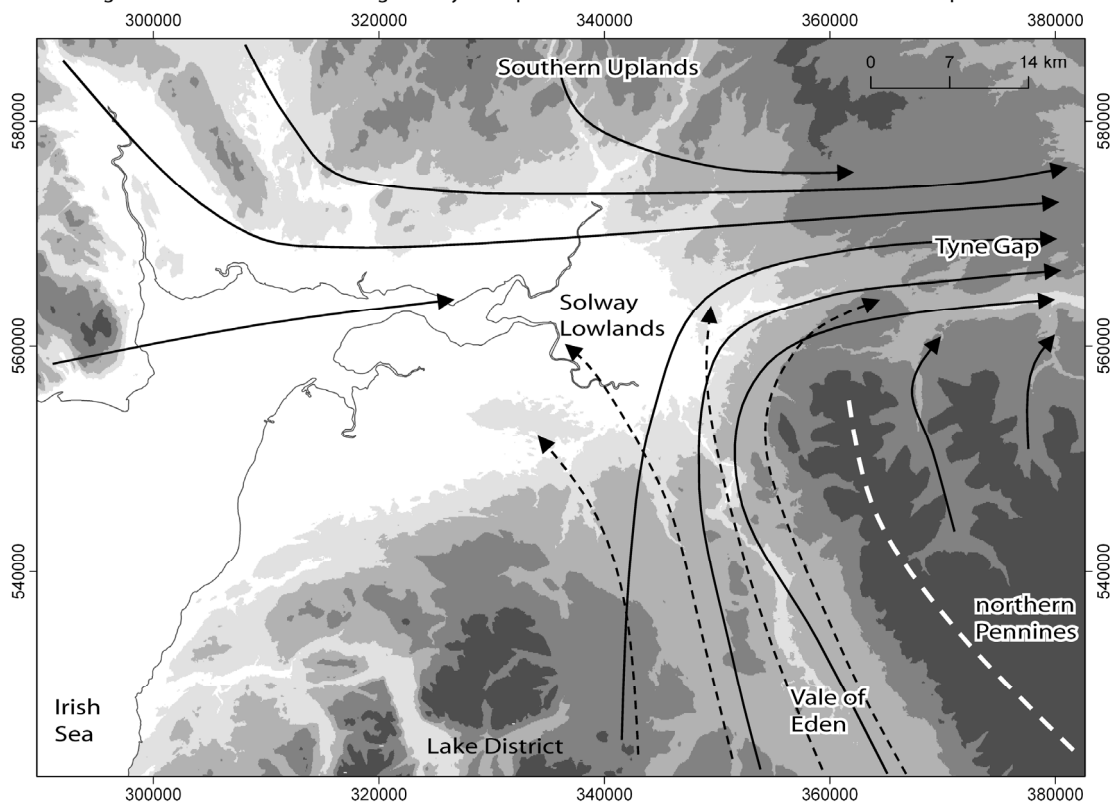
Formation of a glacial-lake in the Solway Lowlands must have occurred when drainage westwards into the Irish Sea Ice Basin became impeded (Fig. 6.14b). The most logical reason for this involves the re-advance of Scottish ice eastwards into the Solway Lowlands, cutting off drainage pathways to the west and north, and causing the natural topographic basin to rapidly fill with meltwater. The sharp contact between the clastic varves and underlying outwash implies that this switch to glaciolacustrine conditions occurred rapidly. Indeed the concept of a re-advance is supported at Blackhall Wood where the varves thicken up-sequence, typical of an increasingly ice-proximal depositional environment (cf. Smith & Ashley, 1985). The slightly thicker varves observed at the very base of the sequence probably relate to initially high sediment fluxes derived from unconsolidated proglacial sediment (cf. Palmer *et al.*, 2008b) or ice-proximity during earlier retreat.

The spatial extent of the lake is hard to estimate due to the likely time-transgressive mode of formation associated with changes in the ice-sheet configuration directly impacting upon the evolution of the ice-contact lake. This is also compounded by both the re-advance of ice westwards into the Irish Sea Basin which would have eroded much of the evidence and the subsequent Scottish re-advance (flow phase SF1: Chapter 2), which produced a separate tripartite division in the northern-most region of the Solway Lowlands (cf. Stone *et al.*, in press). However, the height of the laminated clay-silt lithofacies are generally between 46–54 m O. D. indicating that the lake surface must have been at least 54 m O. D. This is deep enough to flood the entire Solway Basin (an extent > 1,350 km<sup>2</sup>), and enables rough estimates demarcating the topographic boundaries of the lake (Fig. 6.14). The northern and western limits are currently unknown, with the only constraint being the position of the ice-barrier which would have retarded drainage and caused the lake to form. This could have been anywhere from the current Cumbrian coastline to the edge of the Lias plateau (Fig. 6.14b). Borehole evidence of laminated lithofacies indicate that the lake was probably dammed up against the Vale of Eden slope with the lake extent at least 140 km<sup>2</sup> in size (Fig. 6.14). Formation of the lake is therefore regarded as a final stage of ice-free conditions within the Solway Basin. Whereas deposits associated with the ‘main lake’ are consistently observed between 46–54 m O. D., disparate laminated lithofacies observed at a range of higher elevation, up to 133 m O. D. may imply more localised ponding.

## CHAPTER 6: PROGLACIAL LAKE FORMATION IN THE SOLWAY LOWLANDS

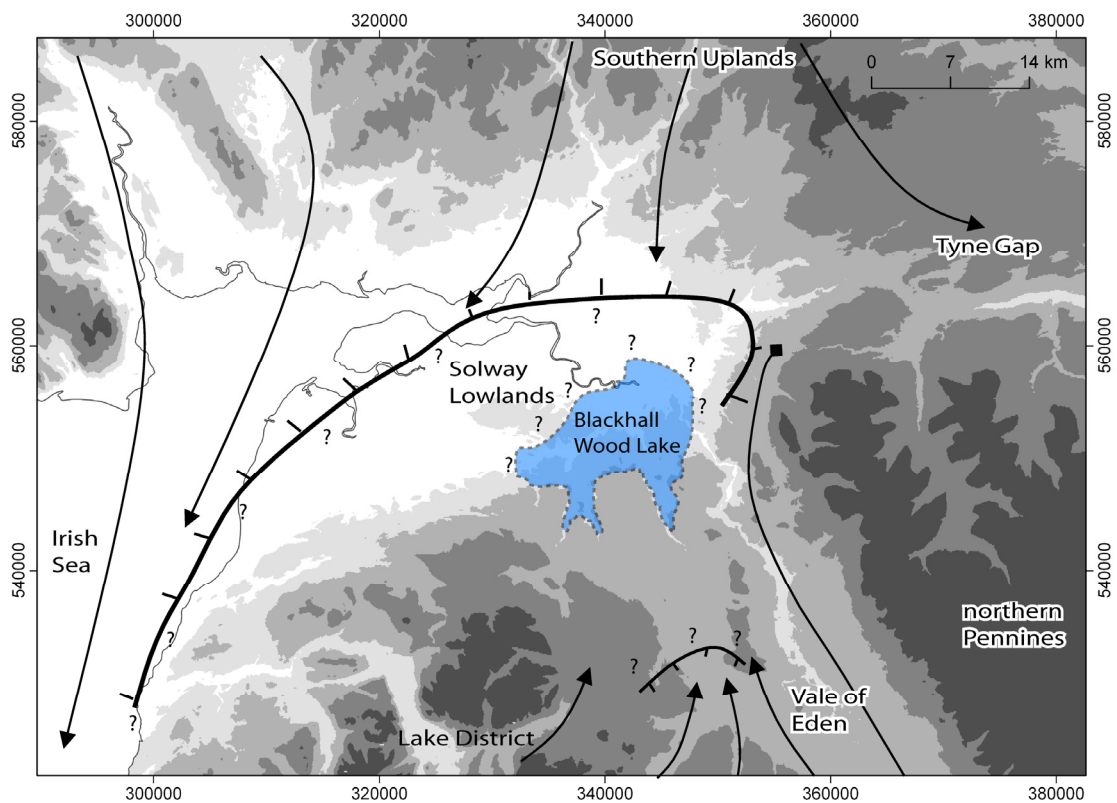
### A Main Glaciation (flow-phases LT1-3)

Convergent ice flow eastwards through the Tyne Gap Ice Stream from Lake District and Southern Upland sources



### B Deglaciation of the Solway Lowlands and formation of Glacial-Lake Blackhall Wood

Development of an ice-free enclave in the Solway Lowlands constrained by an active Irish Sea Ice Stream



## C Blackhall Wood Oscillation (flow-phase LT5)

Re-advance of ice from the Southern Uplands and Lake District into the Irish Sea Basin

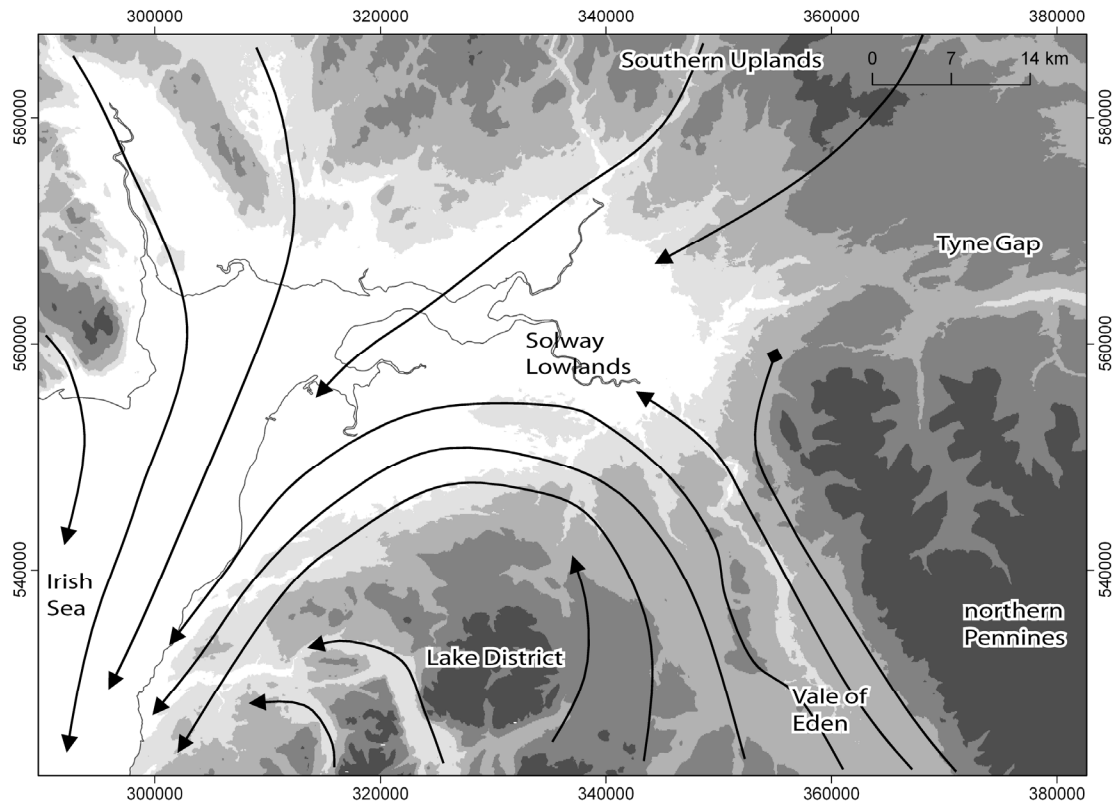


Figure 6.14: Glacial history of the ‘Blackhall Wood’ Oscillation in the Solway Lowlands (for regional correlations see Table 6.3): (a) Prior to the ‘Blackhall Wood’ event, ice from Scotland and the Vale of Eden converged on the Solway Lowlands before moving eastwards through the Tyne Gap as part of a topographic ice stream (flow phases LT1-3: Chapter 2). During ice-recession out of the Tyne Gap ice lobed into the Solway Lowlands (LF 2: Blackhall Wood); (b) Continued retreat of ice out of the Solway Lowlands led to the development of an ice-free enclave, with ice still active in the Irish Sea Basin. Proglacial-Lake ‘Blackhall Wood’ formed during this stage (although the northern, western and eastern extents are unconstrained). The lake is produced purely on the basis of the contour map in Fig. 6.12b, and is likely to have been significantly larger. The ice free enclave has been traced to the western edge of Cumbria and to the border of Scotland. However, again this is a minimum extent; and (c) Ice from both the Lake District and Scotland re-advanced out into the Irish Sea Basin as a tributary of the Irish Sea Ice Stream (flow phase LT5: Chapter 2). (Topographic map (100 m intervals) overlaid onto a NEXTMap DEM, with lighting from a bird’s eye view).

Event	BGS Lithostratigraphic Formations		Stratigraphy (Blackhall Wood)	Geomorphology: flow-sets (Chapter 2)	Regional Correlation	Possible Triggers (?)	Date (cal. ka BP)
	Cumbria	Dumfries-shire					
Scottish Re-advance	Gretna Till Formation			SF1; LT6	Killard Point Stadial	Heinrich Event 1	~ 16.8
Deglaciation (formation of Glacial-Lake Carlisle)	Great Easby Clay Formation	Plumpe Sand and Gravel Formation					
Blackhall Wood Re-advance	Greystoke Till Formation	Chapelknowe Till Formation	LF5	LT5	Gosforth Oscillation; Clogher Head Stadial (?)	19 ka meltwater pulse	~ 19.5
Deglaciation			LFA3; LF4				
Main Glaciation			LF1; LF2	LT1-3	Dimlington Stadial		
Middle Sands							
Early Scottish Advance	Gillcambon Till Formation						Main Late Devensian

< Table 6.3: Updated Event Stratigraphy and regional correlations for the glacial history of the Solway Lowlands (based on work carried out in this Chapter).

The glacial oscillation within the Solway Lowlands is hard to constrain chronologically (see Table 6.3 for the regional Event Stratigraphy). Superimposed glacial landforms, including eskers at Thursby and the Holme St Cuthbert deltaic complex (Huddart, 1970; Livingstone *et al.*, 2008) coupled with the stratigraphic position of the tripartite sequence within Vale of Eden drumlins relating to late-stage flow into the Irish Sea Ice Basin (Livingstone *et al.*, 2008; Chapter 2, Fig. 2.5) indicates that the oscillation must have occurred prior to the Scottish re-advance (~ 16.8 cal ka BP). The Gosforth oscillation observed within the Sellafield District, Cumbria at ~ 19.5 cal ka BP (Merritt & Auton, 2000), prior to the Scottish re-advance (flow phase SF1: Chapter 2) offers a possible regional correlation with the Blackhall Wood re-advance (flow phase LT5: Chapter 2) (Table 6.3). The evidence in this study is consistent with the reconstruction of Merritt & Auton (2000) at Sellafield, with valley ice retreating back into the Lake District while the retreating Irish Sea Ice Stream remained along the coast, thus leading to the formation of proglacial lakes. The subsequent Gosforth re-advance then coalesced with the Irish Sea Ice Stream, leading to extensive drumlinisation (Merritt & Auton, 2000). All these features are noted in the Solway Lowlands.

Given that the Blackhall Wood oscillation occurred during a relatively early stage of deglaciation, prior to the Scottish re-advance, it is likely that the Irish Sea Ice Stream was still quite extensive, stretching down the west Cumbrian coast (Merritt & Auton, 2000) and reaching at least as far south as the Isle of Man (cf. Thomas *et al.*, 2004).

The data from this study indicate that the subsequent flow of ice, flow-phase LT5 (Chapter 2, Fig. 2.5; Table 6.3), sourced in the Vale of Eden/Lake District was in fact a re-advance (Fig. 6.14c). Ice moved down the Vale of Eden, before swinging round the northern edge of the Lake District and coalescing with the Irish Sea Ice Stream (cf. Livingstone *et al.*, 2008). Structural characteristics exhibited within sections at both Swarthy Hill and Maryport suggest drumlinisation of saturated sediment. Glaciofluvial and glaciolacustrine sediments were glaciotectonised and rafted, with evidence for subsequent de-watering. The northern margin of this ice-flow imprint is constrained by a transition into less elongate drumlins and then hummocky terrain and ribbed moraine (Fig. 6.2) (e.g. Ross *et al.*, 2009). This seems to depict the lateral margin of the flow set where inferred ice velocities were significantly reduced (Livingstone *et al.*, 2008; Ross *et al.*, 2009).

The regional-scale of the Blackhall Wood-Gosforth Oscillation (~ 19 cal. ka BP) implies that it was triggered by an external forcing mechanism. This interpretation is given credence by its close chronological association with the 19 cal. ka BP meltwater pulse (Yokoyama *et al.*, 2000; Clark, *et al.*, 2004; Hall *et al.*, 2006). A rapid 10-15 m rise in sea-level related to a large freshwater perturbation is thought to have led directly to an abrupt decrease in the Atlantic meridional overturning circulation and therefore cooling of the NE Atlantic (Hall *et al.*, 2006). The

geomorphological and stratigraphic evidence for the Blackhall Wood-Gosforth Oscillation, presented in this chapter (and also by Merritt & Auton, 2000) therefore provides a possible link between the 19 cal. ka BP meltwater pulse and the BIIS (Table 6.3). This model is in general accord with recent findings in NE Ireland that elucidate two major re-advances during the deglaciation of the Late Devensian Irish Ice Sheet; the Clogher Head Stadial, between 18.3 – 17.0 cal. ka BP, and the Killard Point Stadial after 17.0 cal. ka BP (McCabe, *et al.*, 2005, 2007) (Table 6.3). These offer possible pan-Irish Sea correlations with the Blackhall Wood-Gosforth Oscillation and Scottish Re-advance respectively.

A common theme emerging from the NE sector of the Irish Sea Basin is the repeated development of ice-dammed lakes, ponded-up against local topographic highs in ice-free enclaves along the margins of the Irish Sea Ice Stream. The northern and western borders of the ‘Blackhall Wood’ lake are as yet, unconstrained, and therefore, could possibly have stretched at least 1,350 km<sup>2</sup>, while further ponding dammed by the Irish Sea Ice Stream has been proposed by Merritt and Auton (2000) along the west Cumbrian coast, during the Gosforth Oscillation. Furthermore, lake development seems to have been a regular occurrence throughout the Solway Lowlands, with the Holme St. Cuthbert delta (Huddart, 1970, 1991, 1994), and glaciolacustrine deposits east of Carlisle (Dixon *et al.* 1926; Trotter, 1929) recording the development of large ice-dammed lakes during later-stage episodes of retreat and re-advance. Another significant lake is thought to have existed in the Machars, SW Scotland (Salt, 2001). There, ice-contact sub-aqueous cones at 50 m O. D., stretching from the Isle of Whithorn to Arbrack (Charlesworth, 1926, Salt, 2001), have been interpreted to mark the location of a lake dammed up by Scottish ice to the NE and Lake District ice coming out of the Solway Firth (Salt, 2001). The influence of widespread lake development on ice dynamics in the Irish Sea Ice Basin (see also Ó Cofaigh & Evans, 2001) offers a potential mechanism (through localised lubrication of the ice-bed interface) for triggering fast ice-flow and promoting instability of the ice-margins (Stokes & Clark, 2004).

## 6.6 Conclusions

Evidence presented in this study indicates that a major glacial oscillation, leading to ice free conditions and proglacial lake formation, occurred within the Solway Lowlands prior to the well documented Scottish readvance at ~ 16.8 ka BP. We refer to this event as the Blackhall Wood oscillation and tentatively correlate it with the Gosforth oscillation (Merritt & Auton, 2000) (Table 6.3). It has also been postulated that this regional-scale oscillation was triggered by the 19 cal. ka BP meltwater event (Clark, *et al.*, 2004).

The interpretation of the lamination silt and clay lithofacies within the ‘tripartite sequence’ as varves has allowed a subglacial hypothesis to be discarded in favour of proglacial lake formation.



Micromorphological analysis of the varved sediments has enabled a detailed model of the lake's development to be formalised, and the derivation of a high resolution 'floating point' chronology.

The 'Blackhall Wood oscillation' demonstrates non-linear interplay between ice in the Irish Sea and ice in the Solway Lowlands. Despite the scale of the oscillation which saw one of the core regions of the BIIS become ice free, ice persisted in the Irish Sea Basin, probably as an ice stream. This period of ice-free conditions was then followed by a significant re-advance of ice capable of flowing out into the Irish Sea Basin, depositing thick diamicton sequences and moulding elongate landforms (flow-phase LT5: Chapter 2). This indicates that the oscillation was not a small-scale fluctuation, but rather a period when ice in a core region of the BIIS and in close proximity to key upland dispersal centres rapidly lost and then gained mass, whilst other regions remained relatively stable.

This study suggests that the Irish Sea Ice Stream was still active at a late stage of the Main Late Devensian glaciation, at a time following the 'Blackhall Wood oscillation'. The Solway region itself does not seem to have acted as a major tributary to the Irish Sea Ice Stream barring both the initial contraction and subsequent re-advance of ice associated with the 'Blackhall Wood oscillation'. Prior to this oscillation the geomorphic evidence indicates an easterly ice trajectory through both the Stainmore and Tyne gaps from the Vale of Eden and Solway Lowlands (flow-phases LT1-3: Chapter 2).

### 6.7 Acknowledgments

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## Chapter 7

# Re-advance of Scottish ice into the Solway Lowlands (Cumbria, UK) during the Main Late Devensian deglaciation

### Abstract

The complex glacial geomorphology and stratigraphy of the Solway Lowlands (Cumbria, UK) reflects dynamic ice flow during the Main Late Devensian glaciation, with numerous ice-flow shifts, and re-advances now recognised. The final incursion of Scottish ice into the region (Scottish Re-advance) during a late stage of deglaciation has provoked widespread debate and even scepticism arising from ill-defined marginal limits, and a paucity of landforms, sediments and dates. In an attempt to resolve some of these issues this paper has applied both geomorphological and stratigraphic techniques to critically review evidence pertaining to the Scottish Re-advance. A major deltaic landform-sediment assemblage at Holme St. Cuthbert; a thin, patchy till sheet; and eskers at Thursby and Sowerby Wood verify that Scottish ice re-advanced into the Solway Lowlands at a late stage of deglaciation. The thin, patchy till sheet associated with the re-advance is indicative of a short-lived pulse, with ice thought to have flowed rapidly across water-saturated sediment and into proglacial lakes dammed against higher ground to the east. The short-lived nature of the re-advance coupled with the buffering effects provided by water-saturated sediment at the ice-bed interface meant the re-advance exerted little depositional, erosional or deformational influence. The Holme St. Cuthbert delta, which evolved from a subaqueous outwash fan marks a major still-stand and development of a large ice-contact lake in the vicinity of Wigton. The extent of the glaciation is hard to elucidate, with the till sheet disparately exposed as far east as the rivers Irthing and Cam, although identification of the Blackhall Wood Re-advance till member probably limits its SE extent to Carlisle. Coeval re-advance of Lake District ice is possible given that a final stage lobate movement of ice out of the Vale of Eden and into the Solway Lowlands is now recognised, however there is no direct stratigraphic correlation or dating control.

## 7.1 Introduction

The acknowledgement by palaeoglaciologists of palimpsest ice-flow signatures (e.g. Boulton & Clark, 1990a, b; Punkari, 1993; Clark, 1997, 1999) coupled with a greater availability of higher resolution DEMs has led to a re-appraisal of the British-Irish Ice Sheet (BIIS) during the Late Devensian (Marine Isotope Stage 2) Glaciation. This has resulted in a greater appreciation of the dynamism and sensitivity of the BIIS, specifically recognising major shifts and switches in ice-flow, rapid streaming from the ice sheet core, and multiple oscillations and re-advances of its

margins (Salt & Evans, 2004; Greenwood & Clark, 2008; Livingstone *et al.*, 2008; Evans *et al.*, 2009). The Solway Lowlands in Cumbria (Fig. 7.1) were situated at the heart of the BIIS, and as such, were affected by the complex interplay between multiple upland ice dispersal centres, including the north Pennines, Lake District and Southern Uplands, the migration of ice divides and the impact of the Irish Sea and Tyne Gap ice streams (Livingstone *et al.*, 2008). This complexity is reflected in the geomorphic and stratigraphic record of the region, which has been interpreted in various ways according to the prevalent paradigm of the time (Goodchild, 1875, 1887; Trotter, 1929; Hollingworth, 1931; Huddart, 1971). A number of unanswered palaeoglaciological questions remain outstanding, the most significant of which relates to the Scottish Re-advance, an indistinct and controversial ice-flow phase which is deemed to have impinged onto the Cumbrian coast during a late stage of the Main Late Devensian deglaciation (Trotter, 1922, 1923, 1929; Trotter & Hollingworth, 1932; Huddart, 1970, 1971a, b, 1991, 1994; Huddart & Tooley, 1972; Huddart *et al.*, 1977; Huddart & Clark, 1994).

In light of recent developments in geomorphological mapping, the identification of several re-advances into the Solway Lowlands (Livingstone *et al.*, in press a) and the recognition of multiple ice-flow phases throughout the central sector of the BIIS (Livingstone *et al.*, 2008; Evans, *et al.*, 2009), it now seems pertinent to re-assess the evidence related to the Scottish Re-advance in the Solway Lowlands, and how this relates to reconstructions of the BIIS more broadly. This paper utilises geomorphological mapping and sedimentological and stratigraphic techniques in order to answer a number of research questions, namely: (a) What geomorphological and sedimentological evidence can be correlated to the re-advance of Scottish ice in the Solway Lowlands?; (b) What was the maximum extent of the Scottish Re-advance in the Solway Lowlands?; (c) Was there a concurrent re-advance of ice from the Lake District?; and (d) What are the glaciological implications of the Scottish Re-advance?

## 7.2 Previous research on the Scottish Re-advance

Evidence for a glacial re-advance into the Solway Lowlands was first proposed by Trotter (1922, 1923, 1929). This evidence comprised a thin upper till with Scottish erratics overlying a series of sands and laminated clays (Dixon *et al.*, 1926; Trotter, 1929; Trotter & Hollingworth, 1932). Further evidence in favour of a Scottish Re-advance includes a number of NW-SE trending eskers at Thursby, Cummertree and Gretna, with outwash deltas deposited in association with the eskers at Gretna (Dixon *et al.*, 1926; Charlesworth, 1926; Trotter, 1929) (Fig. 7.2). Although no terminal moraines were recognised, Trotter predicted that the ice-limit reached up to 134 m O. D., stretching as far east as Lanercost, Brampton and Cumwhitton and as far south as Foulbridge and Bolton Low Houses (Figs. 7.1 & 7.2). The thin till facies, lack of terminal moraines, and generally undisturbed sequences were construed as evidence for a short-lived and transient re-advance phase (Trotter,



1929). During and following the maximum extension of the Scottish Re-advance, Trotter (1929) envisaged a number of glacially impounded lakes and overspill channels forming at successively lower levels as the ice front retreated westwards. At the maximum extent, Lake Carlisle and Lake Lyne formed against the reverse slope of the Tyne Gap, with water draining eastwards via the Gilsland meltwater channel (Trotter, 1929). As ice started to retreat westwards lower level lakes Caldew and then Wigton began to develop (Fig. 7.2), with meltwater escaping westwards via overspill channels which connected with, and drained into, successively lower lakes, depositing a series of deltas (Trotter, 1929).

The Scottish Re-advance concept was re-evaluated by Huddart (1970, 1971a, b, 1991, 1994), Huddart & Tooley (1972), Huddart *et al.* (1977), Huddart & Clark (1994) and Huddart & Glasser (2002) with evidence gathered from the Solway Lowlands used to support a more limited phase of ice movement (Fig. 7.2). Huddart (1970) argued that the Scottish ice failed to extend far beyond Carlisle, with much of the stratigraphic evidence provided by Trotter (1922, 1923, 1929) regarded as patchy or open to re-interpretation as debris-flow deposits. Instead the re-advance limits were predominantly defined by esker deposits at Thursby, a thin upper till west of Carlisle and a major glaciofluvial deltaic complex at Holme St. Cuthbert (Huddart, 1970, 1991, 1994) on the Cumbrian coastal fringe (Fig. 7.2). Lake development was acknowledged (Huddart, 1970, 1981, 1991), but attributed to a prior episode of deglaciation, whilst ‘meltwater overspill channels’ such as at Wampool and Wiza Beck (Fig. 7.2) were re-interpreted as subglacial.

Evidence for a Scottish re-advance is not confined to the Solway Lowlands. In SW Scotland Salt and Evans (2004) record a final stage (Flow Stage G of their reconstruction) of radial ice-flow out of the major valleys in the Southern Uplands and a synchronous re-advance surge of Highland ice along the Loch Ryan Basin. The St Bees push moraine situated on the west Cumbrian coast has been assigned to the Scottish Re-advance (Fig. 7.2) by some researchers (e.g. Huddart, 1994; Merritt & Auton, 2000), although its origins remain controversial (cf. Merritt & Auton, 2000; Williams *et al.*, 2001). The Bride Moraine on the Isle of Man has also been tentatively correlated with the Scottish Re-advance, with deposition inferred to be associated with dynamic retreat of the Irish Sea Ice Stream (cf. Thomas *et al.*, 2004), whilst McCabe *et al.* (1998) and McCabe & Clark (1998) have proposed that the Scottish Re-advance correlates with the maximal extent of the Killard Point Stadial Re-advance (~16.8 cal. ka BP). However, these correlations must be treated with caution as there is no *direct* stratigraphic evidence or well constrained chronological control presently available. Furthermore, the growing realisation that the Cumbrian coastal fringe was subject to multiple re-advances (Merritt & Auton, 2000; Livingstone *et al.* in press a) precludes cross-basin correlation of a single event.

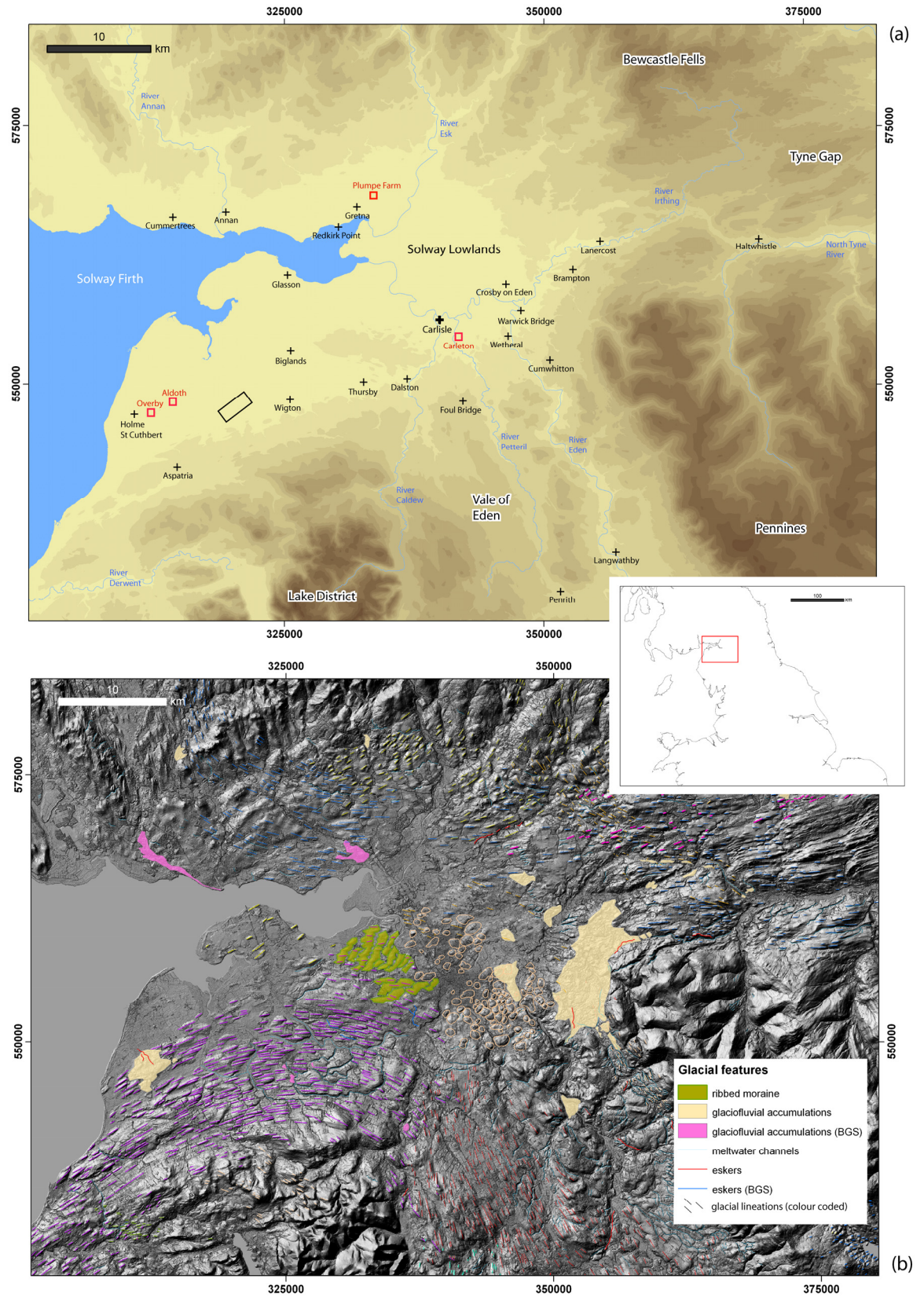
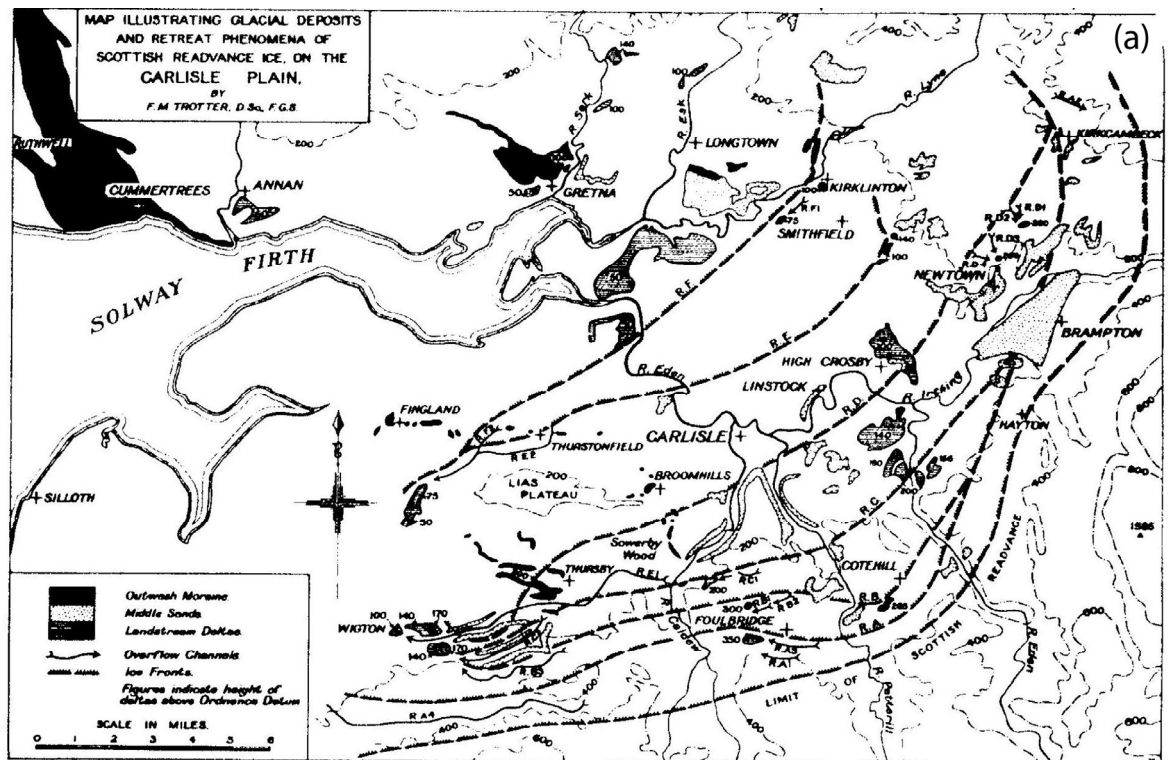


Figure 7.1: (a) Topographic map showing fieldsites (red boxes), borehole data (black box indicates location of boreholes, illustrated in Fig. 7.12) and key locations; and (b) NEXTMap DEM showing the flow-sets and glacial features.

The re-advances that have taken place within the Solway Lowlands include the Scottish Re-advance, and also an earlier 'Blackhall Wood Re-advance' (Livingstone *et al.*, in press a). This re-advance has been correlated with the Gosforth Oscillation (Merritt & Auton, 2000) dated at ~19.5 cal. ka BP, and was characterised by major pro-glacial lake development in an ice-free enclave, followed by re-advance and re-connection with the Irish Sea Ice Stream (flow-phase LT5: Chapter 2; Livingstone *et al.*, in press a).

Due to the rather meagre evidence for a Scottish Re-advance into the Solway Lowlands some researchers have questioned its validity (Pennington, 1978; Evans & Arthurton, 1973; Sissons, 1974; Thomas, 1985). Furthermore, an alternative 'glaciomarine model' proposed by Eyles and McCabe (1989) envisaged rising sea levels into an isostatically depressed Irish Sea Basin causing the rapid retreat of the Irish Ice Stream and resulting in marine limits up to 140 m O. D., thick wedges of ice-contact glaciomarine sediments, raised deltas and mud drapes. This concept largely ignores or re-interprets the evidence for a terrestrial Scottish ice Re-advance in the Solway region, with geomorphic and stratigraphic information such as the Holme St. Cuthbert complex and upper till instead 'forced' into the glaciomarine model (cf. Huddart, 1994).



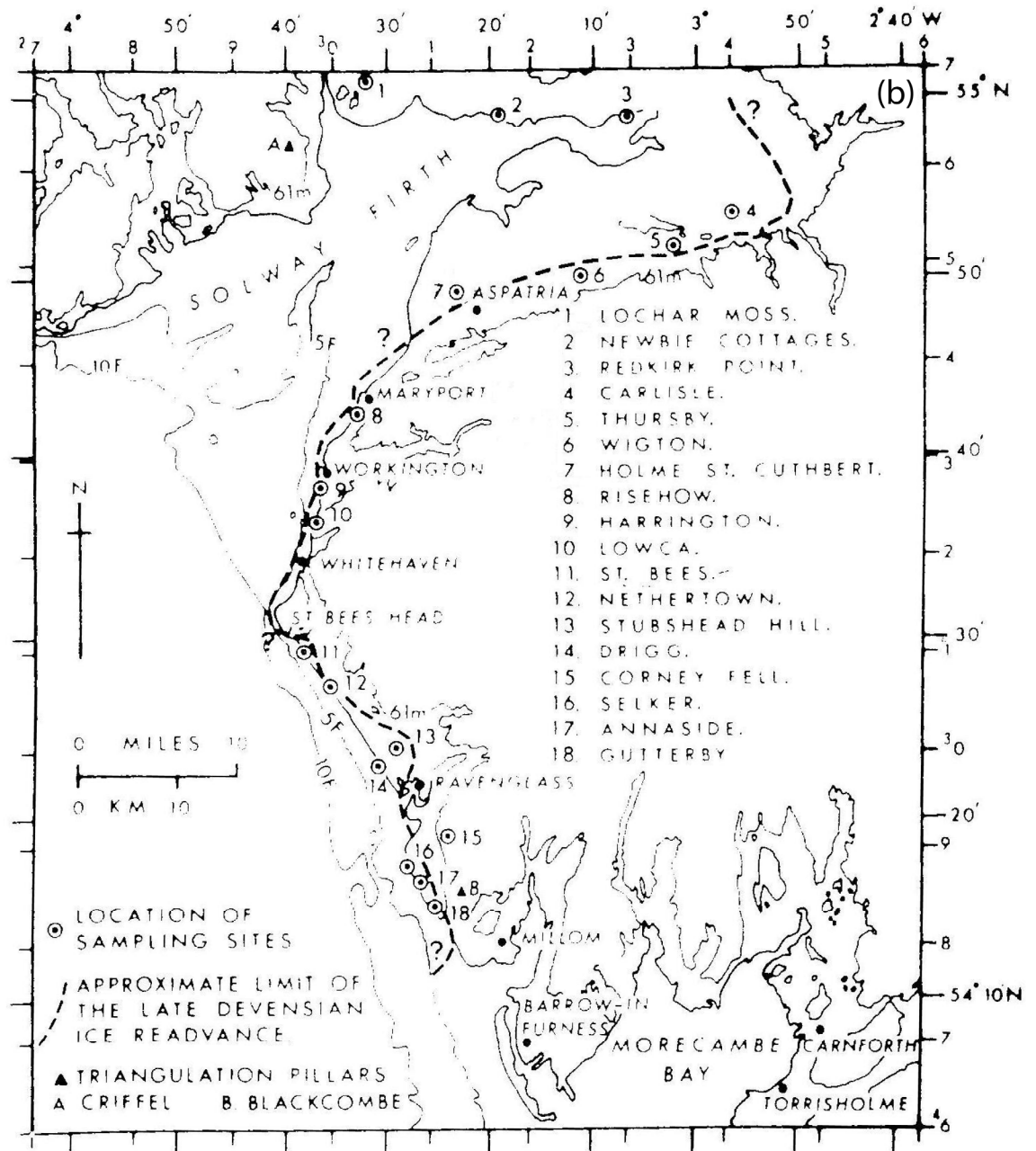


Figure 7.2: Map of the various models of maximum extent for the Scottish re-advance: (a) Trotter, (1929); and (b) Huddart, (1994).

### 7.3 Methods

#### 7.3.1 Glacial geomorphological mapping

Geomorphological mapping involved the recognition and compilation of discrete landform assemblages from NEXTMap digital elevation model (DEM) data. This is a 5 m spatial resolution DEM derived using airborne interferometric synthetic aperture radar (<http://www.neodc.rl.ac.uk/>). In addition, maps of the bedrock geology and superficial deposits (DiGMapGB-625 downloaded



from the BGS) were overlain onto the imagery. Glacial landforms were mapped manually using on screen digitisation. Vectors were used to digitise lineations, meltwater channels and eskers. Polygons were used to digitise hummocky moraine, ribbed moraine, glaciofluvial sediment accumulations and the break of slope exhibited by subglacial lineations (cf. Livingstone *et al.*, 2008). Long profiles along meltwater channels were extracted in order to assess a palaeoglacial versus marginal mode of origin. Lineations were sub-divided into ‘flow-sets’ based on their morphology, length and parallel conformity (Clark, 1997, 1999). This allowed overprinting relationships to be identified and a relative chronology of ice-flow phases to be constructed (Clark, 1999; Livingstone *et al.*, 2008). Ground-truthing was carried out in areas of complexity to verify the mapped glacial landforms, or to substantiate morphological features described in the literature but not observed within the spatial resolution of the DEM. Stoss and lee forms identified from the geomorphological mapping, coupled with published erratic pathways (e.g. Goodchild, 1875; Trotter, 1929; Hollingworth, 1931), were used to interpret flow directions throughout the region.

### 7.3.2 *Sedimentology and stratigraphy*

Field sites pertinent to the Scottish Re-advance were identified throughout the Solway Lowlands (Fig. 1). This included: (a) Overby and Aldoth sand pits in the Holme St. Cuthbert sand and gravel complex; (b) a site on the Scottish border at Plumpe Farm, in an easterly orientated flow-set; and (c) a site exposed at Carleton in hummocky terrain. Texture, sedimentary structure, colour (Munsell colour chart), bed geometry, contacts and inclusions were all measured and logged, from which lithofacies were identified (Evans & Benn, 2004). Lithofacies codes are based upon those of Evans & Benn (2004). Scaled section sketches were drawn at the larger exposures so that the lateral extent of the facies and their architecture could be assessed. Paleocurrent indicators (such as ripples and imbricate gravel), clast macrofabric analysis of the a-axis orientation and dip (Evans & Benn, 2004), and clast lithologies (Bridgland, 1986; Walden, 2004) augmented the sedimentary logging. Borehole logs originally collected by the British Geological Society provided a less detailed but wider coverage, allowing regional stratigraphic correlations.

## 7.4 **Results and Interpretation**

### 7.4.1 *Holme St. Cuthbert sand and gravel complex*

Holme St. Cuthbert (NY 105, 470) is a 9 km<sup>2</sup> spread of sand and gravel situated amongst the strongly lineated glacial terrain of the Cumbrian lowlands (Figs. 7.1 & 7.3). It stretches ~2 km NW-SE between Lowsay Farm and Hards Farm, ~4 km NE-SW between Aldoth and New Cowper, and reaches heights of up to 50 m O. D. (Fig. 3). The complex is superimposed over NE-SW orientated drumlins which comprise a flow set that wraps arcuately around the northern-most edge

of the Lake District and then out into the Irish Sea Basin (Fig. 7.3). The Holme St. Cuthbert complex consists of two distinct morphological assemblages including a flat-topped ridge which runs NE-SW between New Cowper and Round Hill (Fig. 7.3). It is between 0.5 to 1.0 km in width, up to 50 m O. D. and characterised by steep SE and NW facing scarp slopes (Fig. 7.5a). To the west of this scarp is undulating terrain containing a series of depressions and two subdued and winding ridges orientated NW-SE (Fig. 7.3). The genesis of these landforms has been interpreted from their sediment-landform associations using two field sites at Overby and Aldoth sand pits.

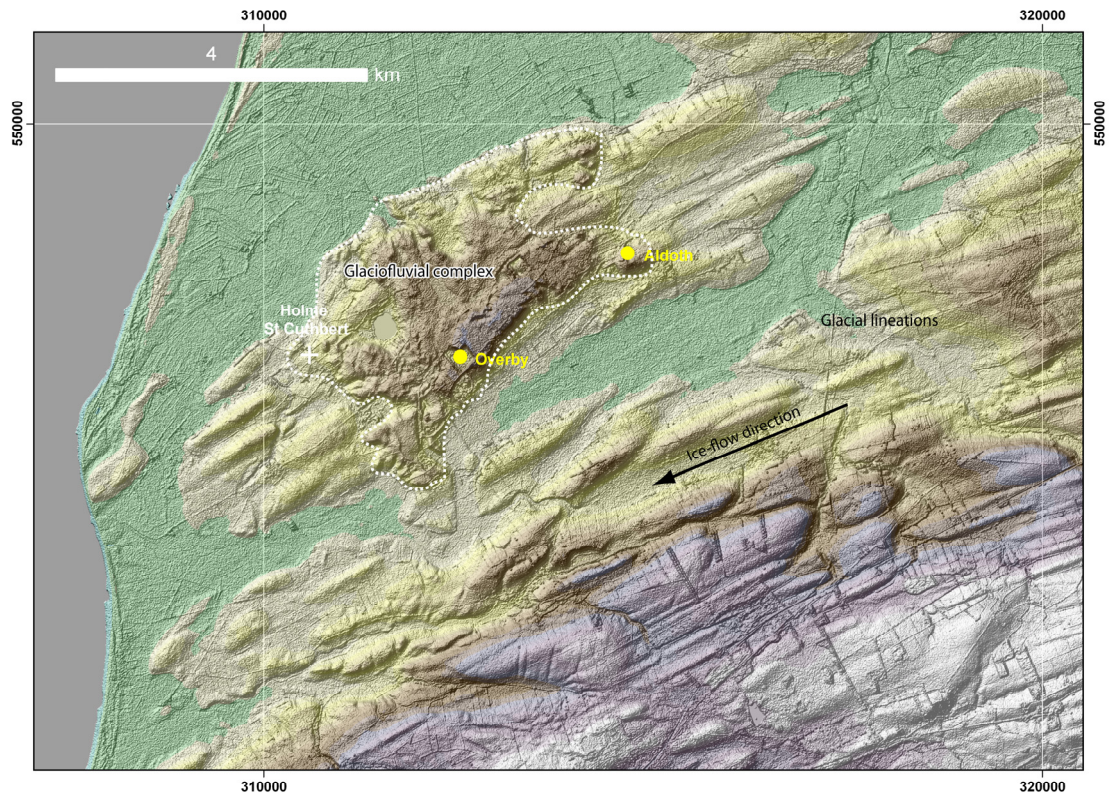


Figure 7.3: NEXTMap DEM of Holme St. Cuthbert glaciofluvial complex superimposed over lineations from flow-phase LT5 (Chapter 2).

Overby sand pit is situated within the flat-topped ridge (NY 125, 471) of the Holme St. Cuthbert sand and gravel complex (Figs. 7.1 & 7.3). The succession has been divided into three main lithofacies associations described and interpreted in turn below, and displayed diagrammatically in Fig. 7.4.

Lithofacies association 1 (LFA OV1) consists of laterally extensive sheets of 0.1 - 1.2 m thick fine to coarse sands which dip gently ( $4^{\circ}$ ) towards the east and are exposed in the bottom 12 - 15 m of the sand pit (Figs. 7.3, 7.6a). This includes horizontally laminated, fine-to-coarse grained sand with occasional granule gravel, thin sheets (0.1 – 0.2) of massive sand, normally graded sand, sinusoidal ripples, type A ripple drift-cross laminations, and type A and B climbing ripples (Allen, 1968, 1973; Jopling & Walker, 1968). The type A and B ripples identified in each of the logged exposures indicate a palaeo-current direction which varies between the east and SSW (Table 7.1).



The lithofacies tend to alternate through horizontal laminations, or massive sand lithofacies, into type A/B climbing ripples (Fig. 7.6a). Sinusoidal ripples and silt/clay bands are infrequently observed, generally at the top of the type A/B climbing ripples. Also identified, commonly towards the top of LFA OV1, were occasional, thin (less than 0.2 m) granule gravel sheets, outsized pebble gravel, and clusters of small (*ca.* 0.5 m thick by 1-2 m wide) trough and lenticular shaped, clast-supported pebble gravels characterised by erosional bases and upwards fining.



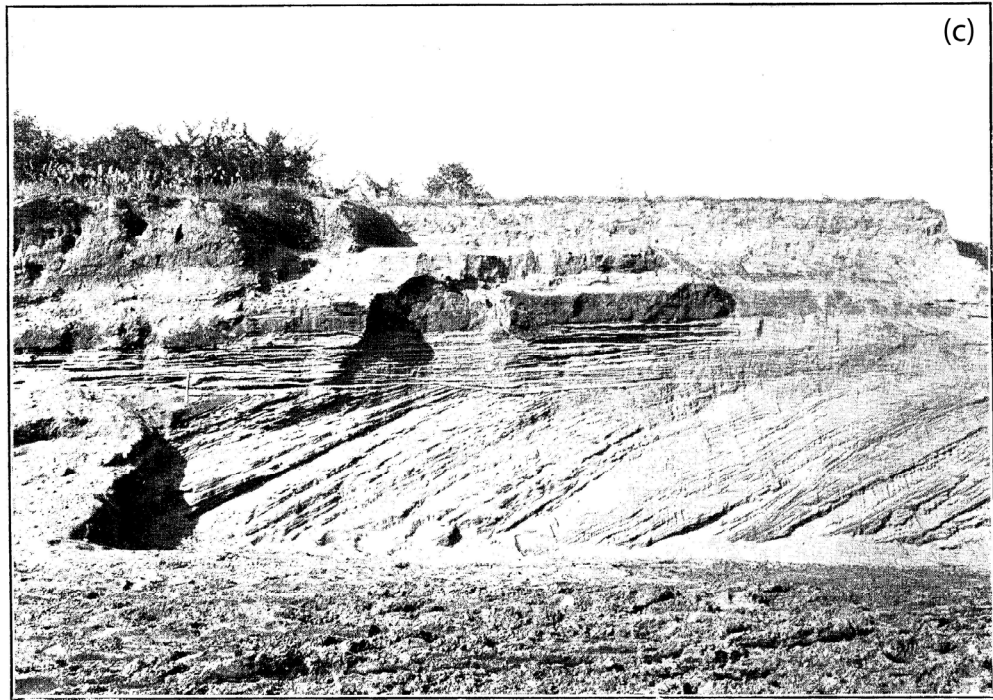
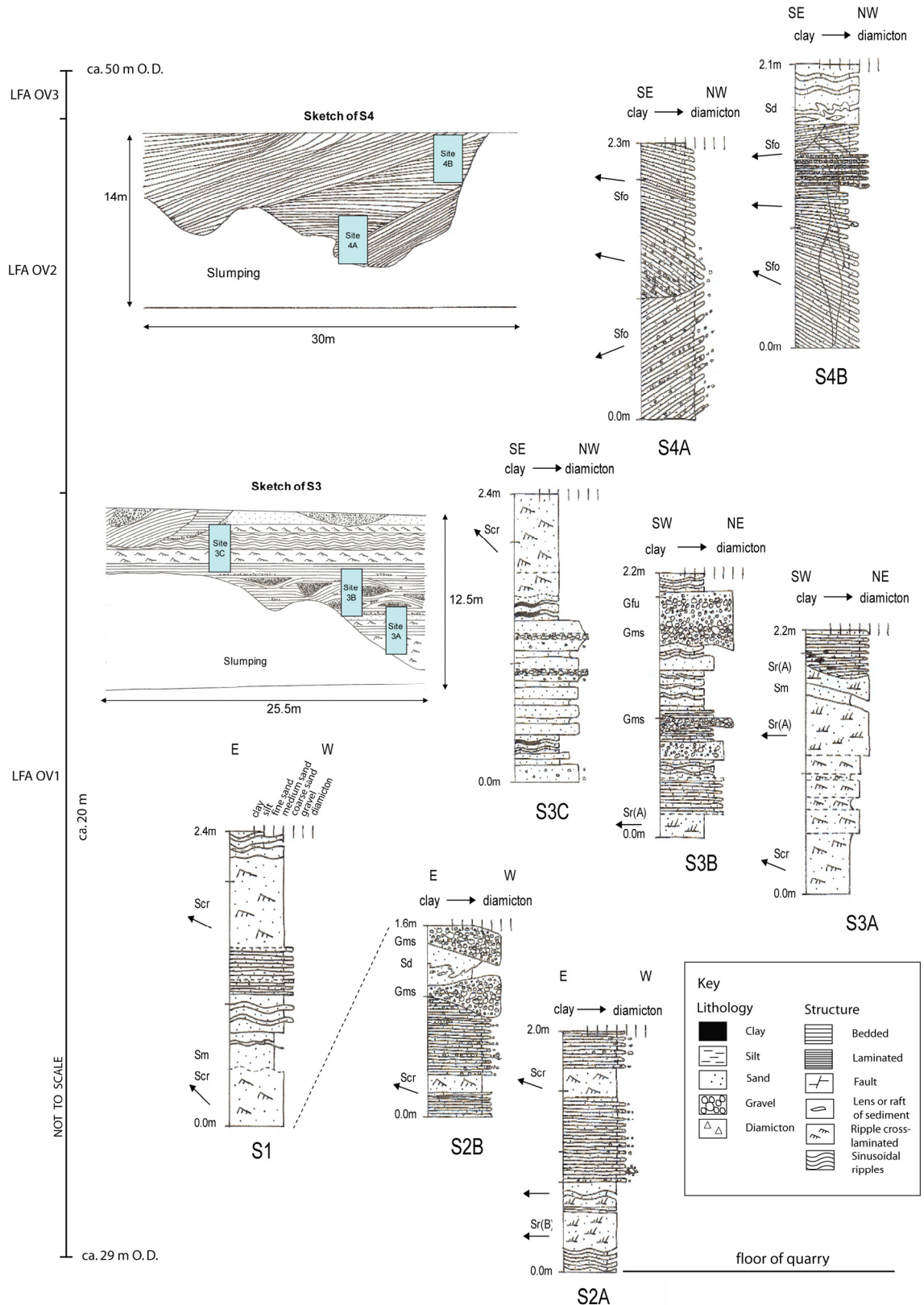


Figure 7.4: Photographs: (a) Looking SW down the ice-contact slope at Holme St. Cuthbert; (b) Subdued esker near Thursby (with sharp bends) – top of esker identified by lighter coloured soil and then framed against trees in background; and (c) Trotter & Hollingworth (1932): picture of delta sedimentology at Warwick Bridge delta (foresets and topsets).

The predominance of fine-grained sediments, including rippled and laminated sand suggests that the deposition of LFA OV1 occurred primarily in a low energy environment. Laminated fine sand and silt/clay bands indicate suspension settling, whilst coarser-grained, medium-to-coarse sand laminae and rippled sand indicate deposition by low energy traction currents or density underflows (Jopling & Walker, 1968; Reineck & Singh, 1975; Smith & Ashley, 1985). Massive and normally graded sand is likely to have been deposited via sediment gravity flows. The complex lithofacies architecture, consisting of cyclical assemblages, is envisaged to have resulted from a fluctuating hydrograph (Smith & Ashley, 1985). This is best exemplified by the progressive changes from coarse/medium sand and granule gravel laminae, through type A/B climbing ripples into sinusoidal ripples which marks a reduction in flow velocity and increase in deposition by suspension fall-out (Jopling & Walker, 1968; Allen, 1973; Smith & Ashley, 1985). The variety of palaeocurrent orientations demonstrates changing influx points, although flow was generally moving south-easterly. Rare scour structures towards the top of LFA OV1 composed of clast-supported normally graded pebble gravel record rapid cut-and-fill processes resulting from turbidity currents and caused by a hydraulic jump (Miall, 1977, 1985; Collinson & Thompson, 1989; Nemec, 1990).

Lithofacies association 2 (LFA OV2) comprises 14 m of steeply dipping planar cross-beds of stratified medium-coarse sand with granule gravel and some out-sized pebble gravel, situated stratigraphically above LFA OV1 in exposures along the entire 60 m NE face of the sand pit (Figs.

7.3 & 7.6b). A series of distinctive units up to 10 m thick are observed within LFA OV2, identified by sharp erosional lower surfaces and dips which range between 6 - 40° towards the SE (with the exception of one unit which dips gently (6°) towards the NW). Some of these units also exhibit slight fining upward sequences from very gravelly coarse sand and granule gravel into medium to coarse sand, with the angle of dip generally decreasing towards the base of the LFA (Fig. 7.6b).



< Figure 7.5: Stratigraphic logs of Overby sand pit.

The dipping, stratified lithofacies of gravelly medium-coarse sand in LFA OV2 are typical of foreset beds formed by cohesionless debris-flows (Smith & Ashley, 1985; Nemec, 1990; Nemec *et al.*, 1999). The presence of distinctive units characterised by internal normal grading, re-activation surfaces and a variety of dips documents changes in the sediment delivery over various timescales either related to changes in the discharge regime or changes in the sediment influx point (e.g. Gustavson *et al.*, 1975; Clemmensen & Houmark-Nielsen, 1981). The general trend of the foreset structures indicates sediment progradation to the SE. The unit dipping in the opposite direction, down to the NW, could be a backset bed formed by very turbulent water (Reineck & Singh, 1975) that cause transient reversals of flow (e.g. Clemmensen & Houmark-Nielsen, 1981; Nemec *et al.*, 1999).

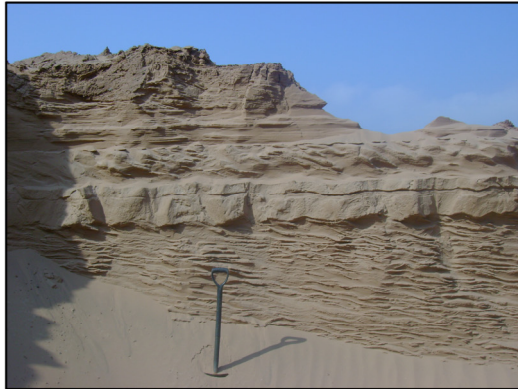
<i>Exposure</i>	<i>Location in section (from bottom of exposure)</i>	<i>Type of ripple</i>	<i>Angle</i>	<i>Palaeo- current</i>	<i>Contact</i>	<i>Texture</i>
2A	0 – 0.5	A	Transverse, sinusoidal, out of phase	W to E	Grades into ripples from sinusoidal structures	Fine-medium sand
	1.47 – 1.67	A	16°	W to E		Fine-medium sand
2B	0.2 – 0.4	A	16°	W to E		Fine-medium sand
I	0 – 0.48	B	36° – 40°	W to E	Grades up into coarse sand	Fine-medium sand
	1.5 – 2.2	A	20° – 24°	W to E	Erosional	Fine-medium sand
3A	0 – 1.15	A	6° – 12°	NNE to SSW	Grades between sand	Grades upwards from fine to coarse sand
	1.15 – 1.65	A	Transverse, sinusoidal, out of phase	NNE to SSW	Erosional	Fine-coarse sand
3B	0 – 0.22	A	Transverse, sinusoidal, out of phase	NNE to SSW	Grades into laminations	Fine-medium sand
3C	1.65 – 1.85	A	20°	NNW to SSE	Grades up from sinusoidal structures	Fine-medium sand
	1.85 – 2.4	B	20°	NNW to SSE	Grades from ripples (1.65 – 1.85)	Fine-medium sand

Table 7.1: Ripples structures.

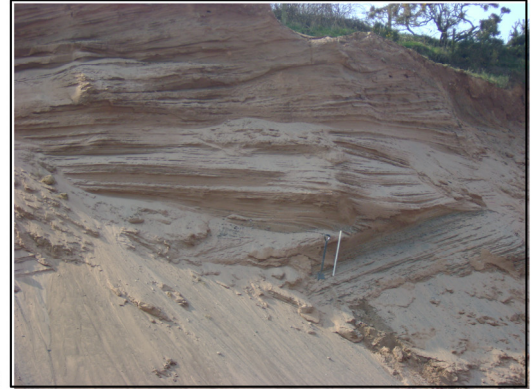
Lithofacies association 3 (LFA OV3) is the capping succession at Overby and is 0.5 – 1.5 m thick and exhibits a series of matrix supported cross-stratified, planar stratified and massive coarse sand and granule gravel lithofacies (Fig. 7.3). Each lithofacies is no more than 0.5 m thick, with erosional lower contacts often related to thin (<0.1 m), discontinuous pebble-gravel lags (Fig.



7.6c). The cross-stratified coarse-sand and granule gravel lithofacies dip south-easterly, in the same direction as the foreset structures of LFA OV2. At the base of LFA OV3 are a number of coarse sand dykes, trending down into LFA OV2, whilst sheared-up blocks of fine-medium sand have also been observed in LFA OV3 (Fig. 7.6c). The top 1 - 2 m of both LFA OV2 and LFA OV3 also show evidence of deformation in occasional small scale convolutions.



(a)



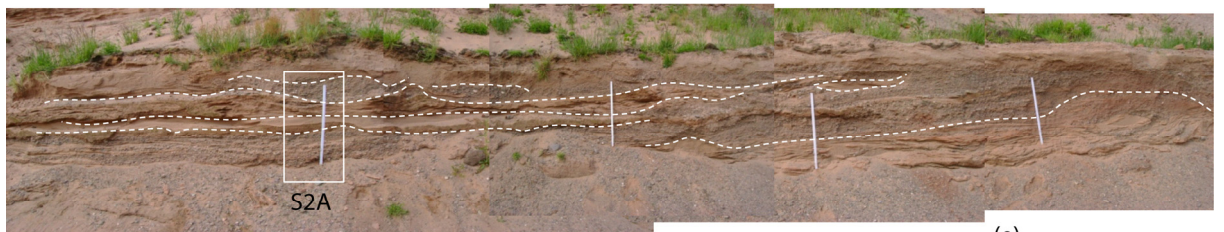
(b)



(c)



(d)



(e)



(f)

< Figure 7.6: Photographs: Overby and Aldoth sand pits: (a) LFA OV1: fine grained deposits (including ripple structures); (b) LFA OV2: foresets; (c) LFA OV3: topsets; (d) LFA A1; (e) Composite of LFA A2; and (f) LFA A3.

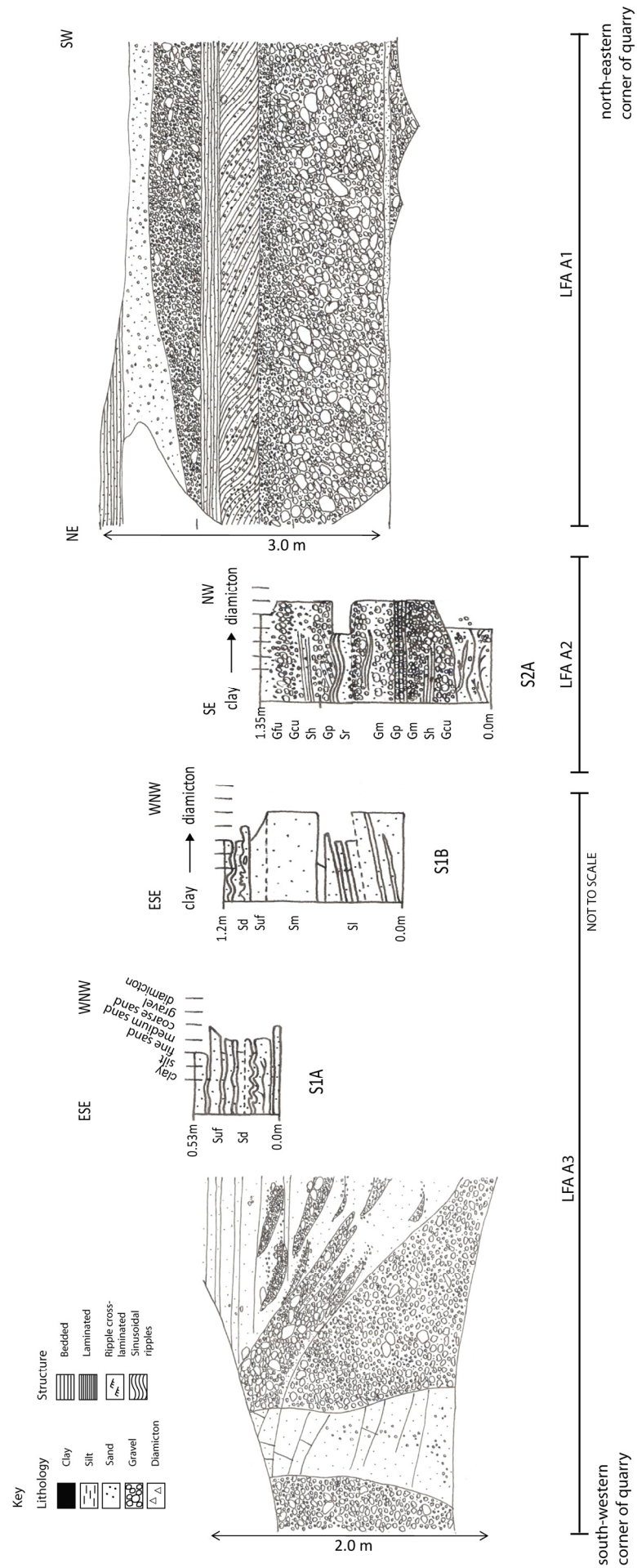
The lithofacies of LFA OV3 are interpreted to be fluvial in origin, with the cross-bedded stratified sand characteristic of dune or bar forms (Miall, 1977, 1985; Smith, 1985; Collinson & Thompson, 1989). The planar stratified coarse sand and granule gravel were deposited by high energy traction carpets (Miall, 1977, Allen, 1984) possibly as bar forms, whilst the massive coarse sand, granule and pebble gravel lithofacies were deposited either during a high energy, high density event (Collinson & Thompson, 1989) or as a debris flow. The pebble gravel lags mark channel floor deposits, with the erosional reactivation surfaces within LFA OV3 demonstrating discrete depositional events, possibly related to a migratory fluvial system (Miall, 1977). The strong erosive contact and marked shift in grain size between LFA OV2 and OV3 indicates a distinct shift in fluvial regime. The dykes of coarse sand are inferred to have been produced by hydrofracturing (e.g. Rijdsdijk *et al.*, 1999; van der Meer *et al.* 1999, 2009), with the downward tapering and infilling from LFA OV3 demonstrating that water escaped and dissipated into LFA OV2.

Aldoth sand pit (NY 148, 483) is situated in the north-eastern corner of the sand and gravel complex in a small sandy hummock (up to 42 m O. D.) which protrudes from the main ridge (Fig. 7.1, 7.3). The succession has been divided into three main lithofacies associations described and interpreted in turn below, and displayed diagrammatically in Fig. 7.7.

Lithofacies association 1 (LFA A1) is 5 - 7 m thick and exclusively found in the north-eastern corner of the sand pit. It is composed of laterally extensive, 0.1 – 1.5 m thick sheets of massive and normally graded, clast-supported gravel lithofacies interbedded with thinner (0.1 – 0.5 m) sheets of horizontally laminated, massive and cross-stratified sand (Figs. 7.6d & 7.7). The sand lithofacies become more prevalent towards the top of the LFA. The gravel is generally pebble sized with some rare cobbles and sub-rounded to rounded clasts (Fig. 7.6d). The beds have sharp contacts and are horizontally organised.

The poorly sorted gravel lithofacies of LFA A1 are thought to have been deposited as cohesionless debris flows (Shanmugam, 2000), whereas the normally graded gravel beds probably formed by high density turbidity currents (Lowe, 1982; Collinson & Thompson, 1989). The interbedded sand lithofacies represent waning flow conditions leading to the development of dune forms (Miall, 1977), traction carpets associated with high density turbidity currents (Miall, 1977, Allen, 1984) and debris flows. Indeed the propensity for high density turbidity currents and debris flows suggest that LFA A1 was deposited in a subaqueous environment (cf. Shanmugam, 2000).





< Figure 7.7: Aldoth sand quarry – stratigraphic logs.

Lithofacies association 2 (LFA A2) was observed towards the southern end of the sand pit, with the approximately 3 m thick sequence consisting of thin ( $< 0.5$  m) sheets of diffusely stratified granule gravel and sand, sinusoidal ripples, infrequently observed and isolated type A ripples, fine sand laminations and occasionally imbricated, normal and inversely graded granule to pebble gravel (Figs. 7.6e & 7.7). The bottom 0.3 m is composed mainly of fine to medium sand, characterised by wavy bedding, ripples and pockets of granule to pebble gravel, which grade up into granule-gravel and gravelly sand-dominated beds (Fig. 7.6e). The beds are generally horizontal to sub-horizontal with both sharp and diffuse boundaries and containing evidence of deformation in the form of centimetre-scale convolutions and sheared, displaced blocks of sand (Fig. 7.6e).

LFA A2 is interpreted as the product of traction deposition related to high and low density turbidity flows in a subaqueous environment (e.g. Lowe, 1982; Shanmugam, 2000). Stratified granule gravel and sand, with evidence of both inverse and normal grading, demonstrates deposition by a traction carpet (Lowe, 1982; Kneller, 1995; Shanmugam, 2000), whilst the sinusoidal ripples and massive sand lithofacies are formed via suspension rain-out (Smith & Ashley, 1985). Imbricate gravel gives a south-easterly palaeo-current, whilst the type A ripples are orientated both towards the SE and the NW. The isolated examples of flow reversal demonstrated by the ripple structures could relate to backsets produced in very turbulent water (Reineck & Singh, 1975). Displaced blocks of sand are attributed to scouring related to a hydraulic jump, followed by rapid deposition (cf. Russell & Arnott, 2003).

Lithofacies association 3 (LFA A3) is composed of bands of massive and normally graded fine to coarse sand, silt and clay, with infrequent examples of trough cross-bedded sand and pebble gravel lithofacies (Figs. 7.6f & 7.7). The bands of sand are sub-horizontally orientated, laterally extensive ( $> 10$  m) and between 0.1 – 0.5 m thick (Fig. 7.6f). They also exhibit fining towards the southern end of the quarry. This LFA, which reaches thicknesses of up to 3.0 m, is observed in the south-western corner of the quarry and overlies LFA A1 and A2 directly. Centimetre-scale convolutions, faults and sheared boundaries in conjunction with a 1.5 m pillar structure composed of heavily faulted sand constitute evidence of deformation.

Massive bands of sand, silt and clay in LFA A3 record deposition by suspension rain-out and density underflows in a low energy environment (Reineck & Singh, 1975; Smith & Ashley, 1985). Normally graded sand bands reflect waning flow conditions, resulting in a loss of competence, whilst massive sand bands reflect rapid deposition due to a loss of flow capacity (Reineck & Singh, 1975). Lateral variations in grain size are attributed to downstream fining. The trough cross-bedded pebble-gravel lithofacies indicate rapidly formed cut-and-fills produced by transient high energy scouring (Miall, 1977, 1985).

The descriptions and interpretations presented above allow a synopsis of the depositional environment at Holme St. Cuthbert. The sand and gravel complex is interpreted as an ice-contact, Gilbert type delta which evolved from a subaqueous grounding line fan (e.g. Powell, 1990; Nemec *et al.*, 1999; Thomas & Chiverrell, 2006). The LFA's of Aldoth sand pit are envisaged to have been deposited as a subaqueous outwash fan (e.g. Rust & Romanelli; Cheel & Rust, 1982; Powell, 1990; Gorrell & Shaw, 1991; Plink-Björklund & Ronnert, 1999; Russell & Arnott, 2003; Winsemann *et al.*, 2007). The exposures are located in a hummock of sand and gravel jutting out from the main body of the delta, with a morphology not dissimilar to an esker-bead (e.g. Bannerjee & McDonald, 1975; Gorrell & Shaw, 1991). Coarse-grained sand and gravel (LFA A1) deposited by traction carpets and cohesionless debris-flows in the zone of flow establishment (ZFE) (Russell & Arnott, 2003) grade rapidly into better sorted and finer grained lithofacies (LFA A2) deposited by episodic turbidity currents (Cheel & Rust, 1982; Postma *et al.*, 1983) and found in the ZFE or zone of transition flow (ZTF) (Russell & Arnott, 2003). LFA A3 documents more distal sedimentation, found in the ZTF or zone of established flow (Russell & Arnott, 2003), with suspension rain-out being the dominant depositional mechanism.

The LFAs at Overby sand pit exhibit all the classical genetic features of a Gilbert-style delta (cf. Smith & Ashley, 1985). The fine grained sediments of LFA OV1 are typical of bottomsets, deposited in a low energy environment (Jopling & Walker, 1968; Gustavson *et al.*, 1975; Cohen, 1979; Clemmensen & Houmark-Nielsen, 1981) distal to the influx point and relating to underflows (ripple structures) and suspension rain-out (clay/silt bands and fine sand laminations). The rhythmicity exhibited by the deposits is a function of fluctuating discharges (Smith & Ashley, 1985), possibly on an annual scale (Gustavson *et al.*, 1975). LFA OV2 displays many of the features of Gilbert-type deltaic foresets, including dipping beds of gravitationally deposited coarse sands and granule gravel, whose bedding becomes shallower lower down in the sequence (Nemec *et al.*, 1999), and distinctive units demonstrating shifts in the influx point (Smith & Ashley, 1985). The fluvial sediments of LFA OV3 are interpreted as topsets, deposited on the delta surface by a migratory river system (Nemec *et al.*, 1999).

An ice-contact origin for the delta has been proposed, based on the following lines of evidence: (a) the undulatory terrain to the NW of the main ridge, which is interpreted as kame and kettle topography composed of glaciofluvial sediments (Cook, 1946; Holmes 1947; Paul 1983) and comprising a series of NW-SE trending eskers and kettle holes; (b) the steep NE-SW orientated ridge which runs along the edge of the kame and kettle topography demarcating the position of the ice-margin; and (c) the formation of a subaqueous fan which must have emanated from a grounding line along the ice front (Powell, 1990).

Palaeo-current indicators, clast lithological analysis and the glacial geomorphology all suggest that ice must have flowed south-easterly out of Scotland and across the Solway Firth. Ripple structures

(Table 7.1) are orientated at a variety of angles ranging between E and SW, whilst the subaqueous outwash fan fines southwards. Clast lithological analysis contains a strong signal from the Southern Uplands (11% Criffell and Dalbeattie granite and 56% Silurian greywacke), whilst the position of the ice-contact slope in relation to the kame and kettle topography indicate that ice must have been situated to the NW of the delta. This interpretation is consistent with the genesis of a glacial lake which must have formed as westerly drainage of meltwater into the Irish Sea Basin and Solway Firth became blocked by ice.

#### 7.4.2 *Regional stratigraphy*

Two sites, Plumpe Farm and Carleton, are pivotal to the regional stratigraphic signature of the Scottish Re-advance. The exposures at Plumpe Farm (NY 334 681) are situated just to the east of the Scottish border amongst subdued and low-lying (<40 m O. D.) W-E orientated glacial lineations (Fig. 7.1). The succession has been divided into two main lithofacies associations described and interpreted in turn below, and displayed diagrammatically in Fig. 7.8. These lithofacies associations are underlain by a lower red-brown diamicton not exposed in the present sections at Plumpe Farm (Phillips *et al.*, 2007).

Lithofacies association 1 (LFA PL1) comprises up to 3.2 m of compact, red fine-medium sand interbedded with a series of silt and clay bands (Figs. 7.8& 7.9a). The sand is generally massive or normally graded, with individual beds up to 1.3 m thick (Fig. 7.9a). Towards the top of the LFA the sediments become locally laminated, exhibiting alternating clay and fine sand bands up to 2 cm thick (Fig. 7.9a). Exposure 2 displays little in the way of deformation, with some shearing between boundaries being evident in the upper half of the LFA. However, exposure 1 displays a series of centimetre-scale deformation structures including normal and reverse faults, sheared contacts and convoluted clay and silt bands (Fig. 7.9b). The extent of deformation is noted to increase upwards through the lithofacies association.

The sediments of LFA PL1 are thought to have been deposited in a glaciolacustrine environment and formed by suspension rain-out (clay/silt bands) and density underflows (sand lithofacies) (Reineck & Singh, 1975). Evidence of deformation, typified by an up-profile increase in shear strain and comprising simple shear and vertical displacement (loading) structures (cf. McCarroll & Rijdsdijk, 2003), suggests that some minor glaciotectionisation occurred (e.g. Hart & Boulton, 1991; van der Wateren, 1995; Evans, 2000). This is mainly concentrated in the upper 0.5 m of the LFA and across the boundary with LFA PL2.

The uppermost lithofacies association (LFA PL2) consists of a reddish (5.0 YR 3/3) matrix-supported diamicton up to 0.9 m thick. The diamicton contains a high frequency of sub-rounded to sub-angular clasts with some striae. The lower contact with LFA PL1 is gradational, with sheared,

fine-grained sands incorporated into the lower 20 cm of the diamicton. Sand and granule gravel pods are also evident within the diamicton, whilst the upper 20 cm is heavily mottled. The principle lithologies are greywacke (59%), siltstone (18%), quartzitic sandstone (7%) and the local Permo-Triassic sandstone (6%). Clast macro-fabric data (Fig. 7.8) reveals a W-E orientation, with an  $S_1$  eigenvalue of 0.59.

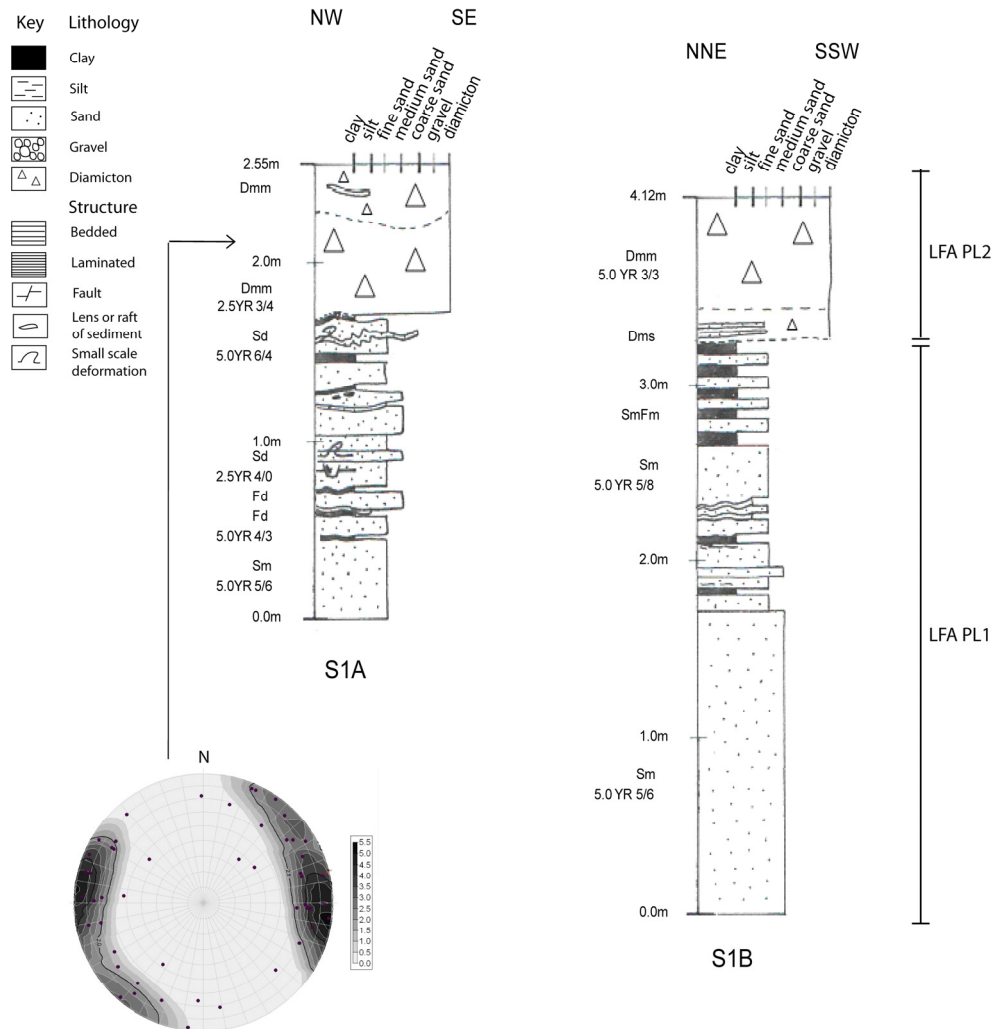


Figure 7.8: Plumpe Farm – stratigraphic logs and clast macrofabric.

LFA PL2 is interpreted as a subglacial traction till (*sensu* Evans *et al.*, 2006) based on the following evidence: (a) clasts that show evidence of striae and contain a far-travelled component derived primarily from the Southern Uplands; (b) the fabric orientation is consistent with the flow-set within which the exposure is situated; (c) the  $S_1$  eigenvalue which indicates a preferred orientation of clasts in a shearing medium (Benn, 1995; Evans, 2000); and (d) the diamicton is underlain by deformed glaciolacustrine sand, silt and clay which has been sheared up and incorporated into the diamicton (e.g. Hicock & Dreimanis, 1989; Benn & Evans, 1996; Hooyer & Iverson, 2000; Evans, *et al.*, 2006).



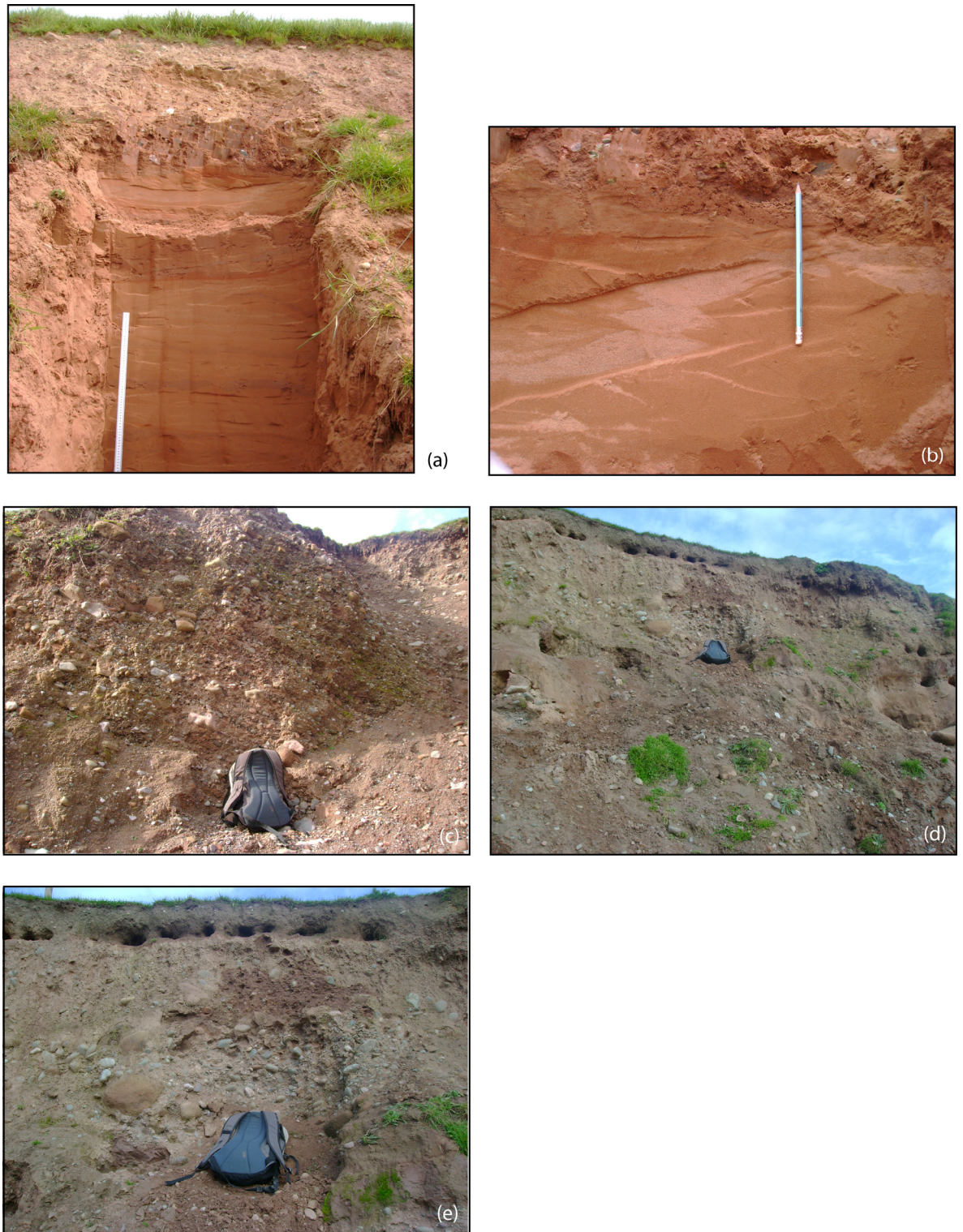


Figure 7.9: Photographs: Plumpe Farm and Carleton; (a) LFA PL1; (b) LFA PL1: deformation; (c) LFA CA1; (d) LFA CA2; and (e) LFA CA3.

Carleton gravel pit (NY 428 527) is located in the central zone of an arcuate tract of hummocky terrain, located in the Petteril Valley to the south east of Carlisle (Fig. 7.1). The mound in which the pit is situated has a weak SW-NE trend, is 0.75 km long by 0.5 km wide and is up to 60 m O. D. (the exposure is situated at 50 m O. D.). The succession has been divided into two main



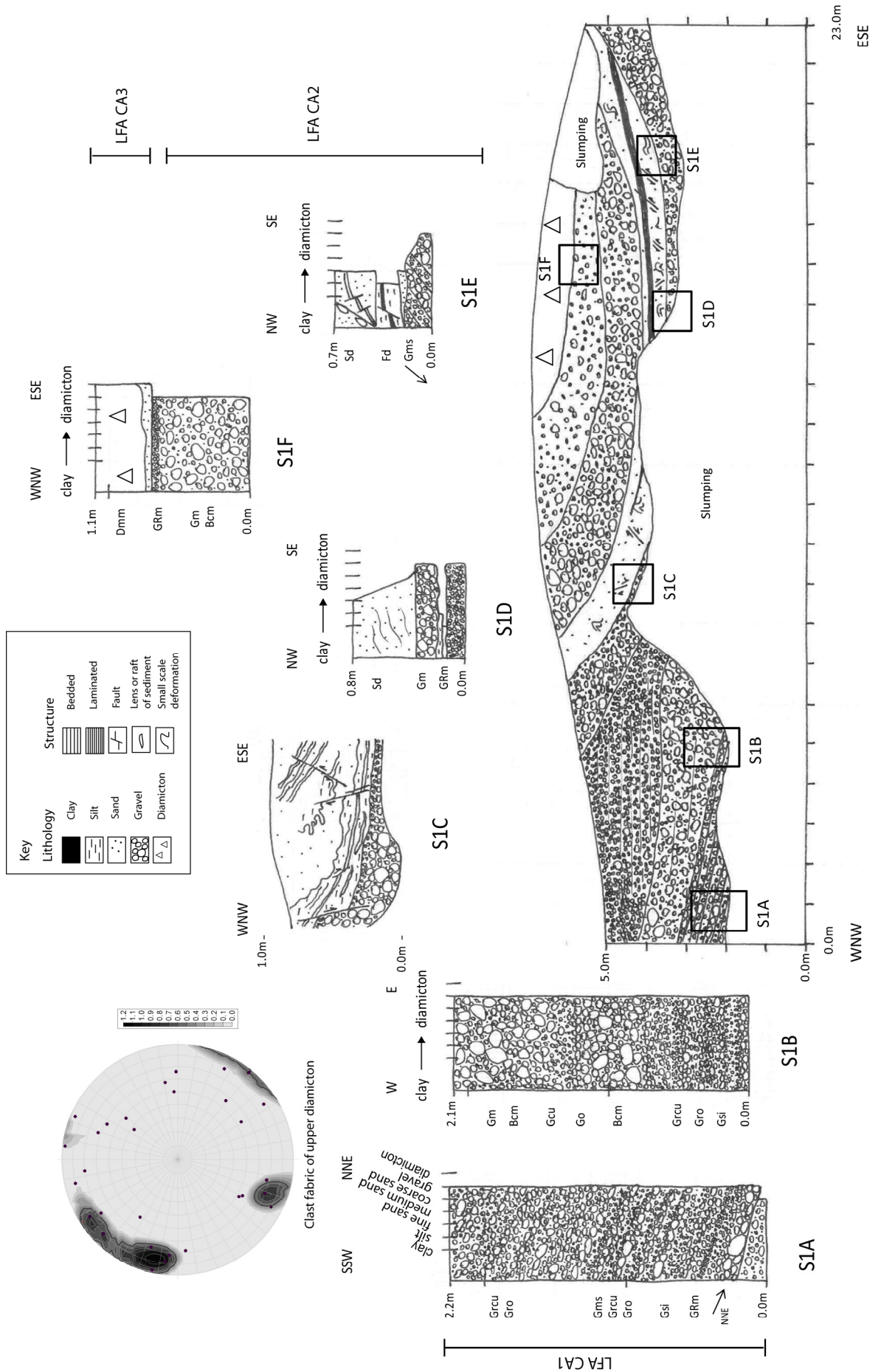
lithofacies associations and one lithofacies described and interpreted in turn below, and displayed diagrammatically in Fig. 7.10.

Lithofacies association 1 (LFA CA1) is composed of NNE dipping ( $22^\circ$ ) foresets containing clast-supported, granule to cobble gravel (max. diameter 30 cm) and coarse sand with some granule gravel (Figs. 7.9c, 7.10). Each foreset bed ranges from 2 – 15 cm in thickness and consists of either a coarsening upwards sequence from coarse sand with granule-gravel to pebble to cobble gravel or alternations of open-work pebble gravel and closed-work granule to pebble gravel (Fig. 7.9c). Rare balls of diamicton are also observed. LFA CA1 is up to 2.2 m thick and is situated in the SSW corner of the gravel pit.

The dipping beds of LFA CA1 are interpreted as deltaic foresets, formed by cohesionless grain flows on a steep slope (Nemec, 1990; Nemec *et al.*, 1999). The inverse grading is produced by dispersive pressures (Bagnold, 1956; Lowe, 1982), whilst openwork pebble to cobble gravels result from clogging in the upper layers by fines during fluctuating discharges (cf. Smith, 1985). The foreset orientations suggest that water flowed NNE, whilst the presence of diamicton balls indicate that the delta must have been either in contact with, or proximal to, the ice front.

LFA CA1 is overlain by a synclinal, 3.5 m thick by 15 m wide LFA (CA2) composed of interbedded clast-supported gravel lithofacies (up to 1.0 m thick), occasional thin (< 5 cm) clay/silt lithofacies and a 0.7 m bed of clay (Fig. 7.9d). The gravels, which dominate the LFA, range from pebble to cobble sized, and are either massive or display coarsening followed by fining upward sequences. The bed of fine-grained sediments is composed of a lower (*ca.* 0.3 m) band of silt capped by massive and normally graded fine sand with some sinusoidal ripple structures and a clay band. This fine-grained unit is heavily disturbed, with evidence of shearing, silt and clay stringers, flame structures, convoluted bands of clay, centimetre-scale reverse faults and a series of normal faults, with offsets of up to 0.3 m, which extend into the gravel lithofacies (above and below). The principle lithologies of LFA CA1 & CA2 are quartzitic sandstone (26%), greywacke (24%), Carboniferous Limestone (13%), Permo-Triassic sandstone (8%) and andesite, rhyolite & basalt (8%).

The laterally extensive, horizontally bounded and clast-supported gravel lithofacies in LFA CA2 are interpreted as longitudinal bars (Boothroyd & Ashley, 1975; Rust, 1975; Miall, 1977, 1985; Smith, 1985; Marren, 2001) formed by tractional deposition in a glaciofluvial environment. The range of coarse-grained material, from granule to cobble size, suggests that there were significant fluctuations in discharge, with very high energies required to move the largest cobbles. The fine-grained sediments are typical of abandonment, or low-flow conditions, leading to the rapid deposition of the suspended sediment load (Miall, 1977; Smith, 1985).



< Figure 7.10: Carleton – stratigraphic logs and clast macrofabric.

At the very top of the succession, LF CA3 is a thin (0.4 m), laterally discontinuous dull reddish-brown (5.0 YR 4/3) diamicton (Fig. 7.9e). The diamicton is massive, clast-rich, with rounded to sub-rounded clasts, and grades via sandy diamicton and granule gravel into the underlying LFA. No striations were observed on clast surfaces. A macro-fabric taken from the diamicton reveals a NW-SE orientation (Fig. 7.10) with an  $S_1$  eigenvalue of 0.46. The principle lithologies are andesite, rhyolite & basalt (31%), greywacke (30%), dolerite (12%), quartzitic sandstone (5%) and mudstone (4%). The diamicton of LF CA3 is characterised by a low  $S_1$  eigenvalue, lacks striated clasts, is discontinuous in nature, has a thin geometry and contains predominantly rounded to sub-rounded clasts typical of subaerial debris-flow deposit (e.g. Lawson, 1981, 1982).

From the above descriptions and interpretations we can make a synopsis of the environment of deposition at Carleton and Plumpe Farm. The exposure at Plumpe Farm forms part of a tripartite sequence consisting of an upper and lower till separated by fine-grained glaciofluvial or glaciolacustrine sediments. The deposition of an upper till which partially glaciotectonised the underlying sand, silt and clay lithofacies implies that ice underwent a re-advance phase. Macrofabric and clast provenance data support the interpretation of a Scottish re-advance emanating from Dumfries-shire and flowing eastwards into the Solway Basin (e.g. Dixon *et al.*, 1926; Trotter, 1929; Phillips *et al.*, 2007). The lack of any major glaciotectonic structures implies that conditions at the ice-bed interface acted to reduce the transmission of shear strain down into the sequence, possibly due to a water lubricated surface (cf. Phillips *et al.*, 2007).

The gravel and sand lithofacies at Carleton demonstrate deltaic and glaciofluvial deposition (e.g. Marren, 2001). Extensive deformation, including normal faults, sheared boundaries, convolutions and flame structures, is characteristic of high water pressures and vertical collapse typically found in a proglacial environment (McDonald & Shilts, 1975; McCarroll & Rijdsdijk, 2003), whilst the presence of diamicton balls support this inference. The clast provenance and palaeo-current indicators suggest that ice was situated to the S/SW of Carleton, with the gravels mainly derived from the local bedrock. The upper diamicton which caps the sequence (LF CA3) is interpreted on a local-scale as a subaerial debris-flow. Geochemical analysis indicates a similar provenance to the upper till of Plumpe Farm (see Chapter 8 for a detailed overview), which when combined with its stratigraphic position, indicates deposition during the Scottish Re-advance. The widespread deposition of this thin capping diamicton throughout the Solway Lowlands (cf. Trotter, 1929; Huddart, 1970) is more akin to subglacial genesis as a discontinuous till sheet. Evidence of debris-flows both here and throughout the Solway Lowlands (Huddart, 1970) probably relate to subsequent mass movements of the till sheet following retreat of the ice.

### 7.4.3 Regional glacial geomorphology

Glacial geomorphological mapping within the Solway Lowlands has identified a series of lineations, meltwater channels, eskers, glaciofluvial deposits, hummocky terrain and ribbed moraine and has facilitated the grouping of lineations into discrete flow-sets (Fig. 7.1). These flow-sets have been used to construct a relative chronology based on identifiable cross-cutting relationships between glacial features (cf. Livingstone *et al.*, 2008 and Chapter 2). This has allowed a number of late-stage ice flows to be identified, which could relate to a Scottish Re-advance.

Livingstone and co-authors (in press a) assigned the arcuate belt of lineations, hummocky terrain, and ribbed moraine of the Solway Lowlands, along with the NE-SW orientated drumlins located in Bewcastle Fells and at Glasson (Fig. 7.1), to the “Blackhall Wood Re-advance” (flow-phase LT5: Chapter 2, Fig. 2.5). This re-advance flowed south-westwards into the Irish Sea Ice Basin as a tributary of the Irish Sea Ice Stream (Roberts, *et al.*, 2007; Livingstone *et al.*, in press a). The Blackhall Wood Re-advance phase is subsequently cross-cut by the Holme-St. Cuthbert deltaic sequence (flow-phase SF1: Chapter 2, Fig. 2.5) (Fig. 7.1), which has been assigned to the Scottish Re-advance (Huddart, 1970, 1981, 1991), and also a northerly orientated set of lineations moving out of the Vale of Eden and into the Solway Lowlands (Fig. 7.1), which must relate to a late-stage oscillation of Lake District ice (flow-phase LT6: Chapter 2, Fig. 2.5). A thick band of diamicton towards the margin of this lobate re-advance (cf. Livingstone *et al.*, in press a) could mark the position of an associated moraine.

The W-E orientated flow-set of lineations marking the flow of ice down the Annan Valley and into the Solway Lowlands (Fig. 7.1) has been correlated with the similarly orientated Tyne Gap flow-set which demarcates the easterly movement of ice across the Pennines (cf. Livingstone *et al.*, 2008, in press b). However, the stratigraphic sequence at Plumpe Farm (Fig. 7.8) suggests a similarly-orientated Scottish Re-advance which deposited a thin till sheet across the region around Gretna. This may or may not have acted to remould or partially remould the Tyne Gap flow-set leading into the Solway Lowlands.

Many of the meltwater channels incised into lineations formed during the ‘Blackhall Wood Re-advance’ (Livingstone *et al.*, in press a; Figs. 7.1 & 7.11) have caused controversy. This is especially true for the Wiza Beck meltwater channels which have been variously interpreted as ice-marginal channels associated with the Scottish Re-advance (Dixon *et al.*, 1926; Trotter, 1929) and subglacial channels (Arthurton & Wadge, 1981; Huddart, 1970). Huddart (1970) suggests that these channels formed prior to the Scottish Re-advance, during ice recession. Similarly the Wampool meltwater channel has been associated both with the retreat of the ‘Main Glaciation’ (Huddart, 1970) and the subsequent Scottish Re-advance (Trotter, 1929; Trotter & Hollingworth, 1932).

Cutting across the western corner of the Lias plateau around Biglands (Fig. 7.1) are NE-SW orientated channels approximately 4 km long. The channels have slight up and down profiles (about 2 m amplitude), which dip gently down to the SW, and are below 30 m O. D. (Fig. 7.11)

Wampool meltwater channel is a broad-bottomed channel (up to 0.5 km wide) over 11 km in length that trends westwards from the River Caldew out towards the Irish Sea coast (Fig. 7.1). The channel geometry is 'U' shaped, with a gentle long profile (Fig. 7.11) descending westwards from 40 to 18 m O. D. A number of the Wiza Beck channels are confluent with Wampool's southern bank (Fig. 7.1).

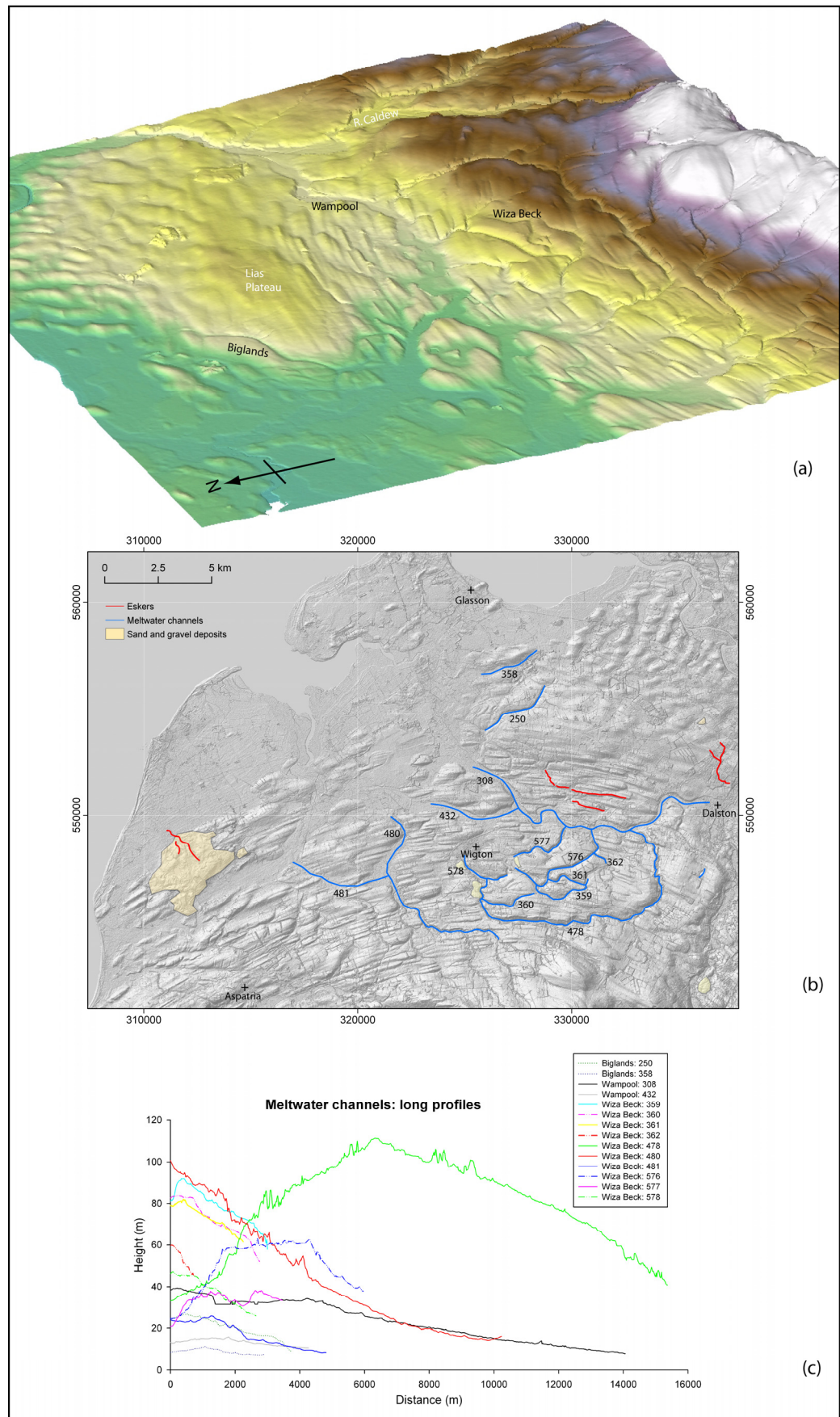
Wiza Beck meltwater channels, on the northern-edge of the Lake District, form an arcuate network trending southwards from Wampool meltwater channel and then westwards towards the coast (Fig. 7.1). The channels are broadly aligned and are characterised by undulatory long profiles which are deeply incised into the underlying drumlins (Fig. 7.11). The arcuate channels dip steeply towards Wampool meltwater channel, and gently W-E down towards the coast. Channel lengths range between 1 and 16 km, with heights of between 10 – 110 m O. D. (Fig. 7.11).

Despite being below the spatial resolution of the NEXTMap DEM, ground-truthing has confirmed the existence (cf. Dixon *et al.*, 1926; Huddart, 1970, 1981) of a number of subdued (generally <5 m tall) eskers. These are correlated with the Scottish re-advance because they are superimposed on lineations associated with the earlier Blackhall Wood re-advance (Figs. 7.1, 7.5). At Thursby, two discontinuous eskers were observed, running W-E and with a series of right-angled kinks (Figs. 7.1 & 7.5). Another esker at Sowerby Wood, just north of Dalston (Fig. 7.1) on the Lias plateau is orientated SE-NW and branches northwards into two strands. These eskers therefore provide strong evidence for the re-advance of Scottish ice into the Solway Lowlands as far east as Sowerby Wood.

A number of flat-topped hills have been mapped to the east of Carlisle (Fig. 7.1). The two most distinctive flat-topped hills overlap at Wetheral (NY 468, 548) and Warwick-Bridge (NY, 455, 563), with the former being 1.8 x 0.8 km in dimension and 54 m O. D. and the latter being 2.0 x 1.0 km and 43 m O. D. (Fig. 7.1). Two further flat-topped hills have been recorded at Crosby-on-Eden (NY 458, 602) and Cam Beck (NY 499, 620), which are 30 m and 60 m O. D. respectively (Fig. 7.1). These have been identified as deltas due to their morphology (Fig. 7.1) and sedimentology (borehole data (BGS); Trotter & Hollingworth, 1932, Fig. 7.5), plus the presence of fine-grained lacustrine deposits throughout the region (Jackson, 1979; BGS borehole data).

These deltaic complexes must therefore record the evolution of a major glacial lake, formed against the reverse slope of the Tyne Gap, with meltwater drainage to the west blocked by an ice mass situated in the Solway Lowlands. The various levels of the flat-topped hills indicate a general trend of falling water levels westwards towards Carlisle. This glacial lake 'Glacial-Lake Carlisle' has been previously identified, although the timing of its formation is controversial. It could either have

formed during the ‘Scottish Re-advance’ (Trotter, 1929), or during the main phase of deglaciation (Huddart, 1970, 1981, 1991).





< Figure 7.11: (a) 3D NEXTMap DEM of the glacial meltwater channels of Biglands, Wiza Beck and Wampool incised into an arcuate flowset trending round the northern margin of the Lake District; (b) Plan-view of the Wiza Beck, Wampool and Biglands meltwater channels (NEXTMap DSM) (numbers refer to long-profiles on Fig. 7.11c; and(c) meltwater channel long-profiles. Biglands meltwater channels (long profiles taken NW-SE); Wiza Beck/Wampool meltwater channels (long profiles taken W-E).

## 7.5 Discussion

The following discussion aims to answer the four research questions posed at the beginning of this paper based on a critical review of previous research regarding the Scottish re-advance and our new evidence. The primary objective is to ascertain which landforms and sediment assemblages can be confidently correlated with a possible Scottish Re-advance given that we now understand that an earlier ('Blackhall Wood') Re-advance occurred in the central sector of the BIIS. A thorough examination of the evidence pertaining to the 'Scottish Re-advance' is crucial if its configuration, impact and behaviour during the deglaciation of the BIIS are to be resolved.

### 7.5.1 *What geomorphological and sedimentological evidence can be correlated to the re-advance of Scottish ice?*

The Holme St Cuthbert ice-contact deltaic complex provides compelling evidence for a Scottish Re-advance subsequent to the Blackhall Wood Oscillation (Livingstone *et al.*, in press a), during a late stage of deglaciation. A glacial lake developed when ice situated along the Cumbrian coast impounded meltwater draining westwards into the Irish Sea Basin. The maximum extent of the lake is difficult to delimit, although borehole evidence (BGS) of infrequently observed thin clay facies capping the sequence have been identified between Aspatria and Wigton (Fig. 7.12). Despite this sporadic sedimentological evidence and absence of observable strand-lines, the height of the foreset structures, identified at Overby sand pit at *ca.* 49 m O. D., indicate that the lake could potentially have extended over much of the Solway Lowlands. The lack of observable evidence for the lakes existence, apart from the delta at Home St. Cuthbert suggests that it was a short-lived event.

Plumpe Farm offers the best stratigraphic evidence for a re-advance of Scottish ice into the Solway Lowlands. The tripartite sequence clearly demonstrates re-advance of ice over glaciolacustrine sediments from a Scottish sourced ice-flow. This conclusion is in agreement with work carried out by Dixon *et al.* (1926), Trotter (1929), Phillips *et al.* (2007) and Stone *et al.* (in press). They observed a number of exposures within the Gretna region, extending west to as far as Redkirk point (Fig. 7.1), characterised by tripartite divisions (Dixon *et al.*, 1926; Trotter 1929). Clast macrofabrics and provenance data demonstrate that the till was deposited by an easterly flowing ice mass (also see Phillips *et al.*, 2007). The diamicton (LF CA3) at Carleton is correlated with the till

at Plumpe Farm, and is envisaged to form part of an extensive till sheet which stretched as far east as Lanercost (Trotter, 1929; Trotter & Hollingworth, 1932b). Huddart (1970) argued that the discontinuous nature and thin geometry of the deposits was indicative of subaerial debris flows. However, the lateral extent of deposition ( $>15$  km) is hard to reconcile with this interpretation, while tills are often thin ( $>2$  m) and patchy (e.g. Rooney *et al.*, 1987) and should therefore not preclude a subglacial genesis (cf. Evans *et al.*, 2006).

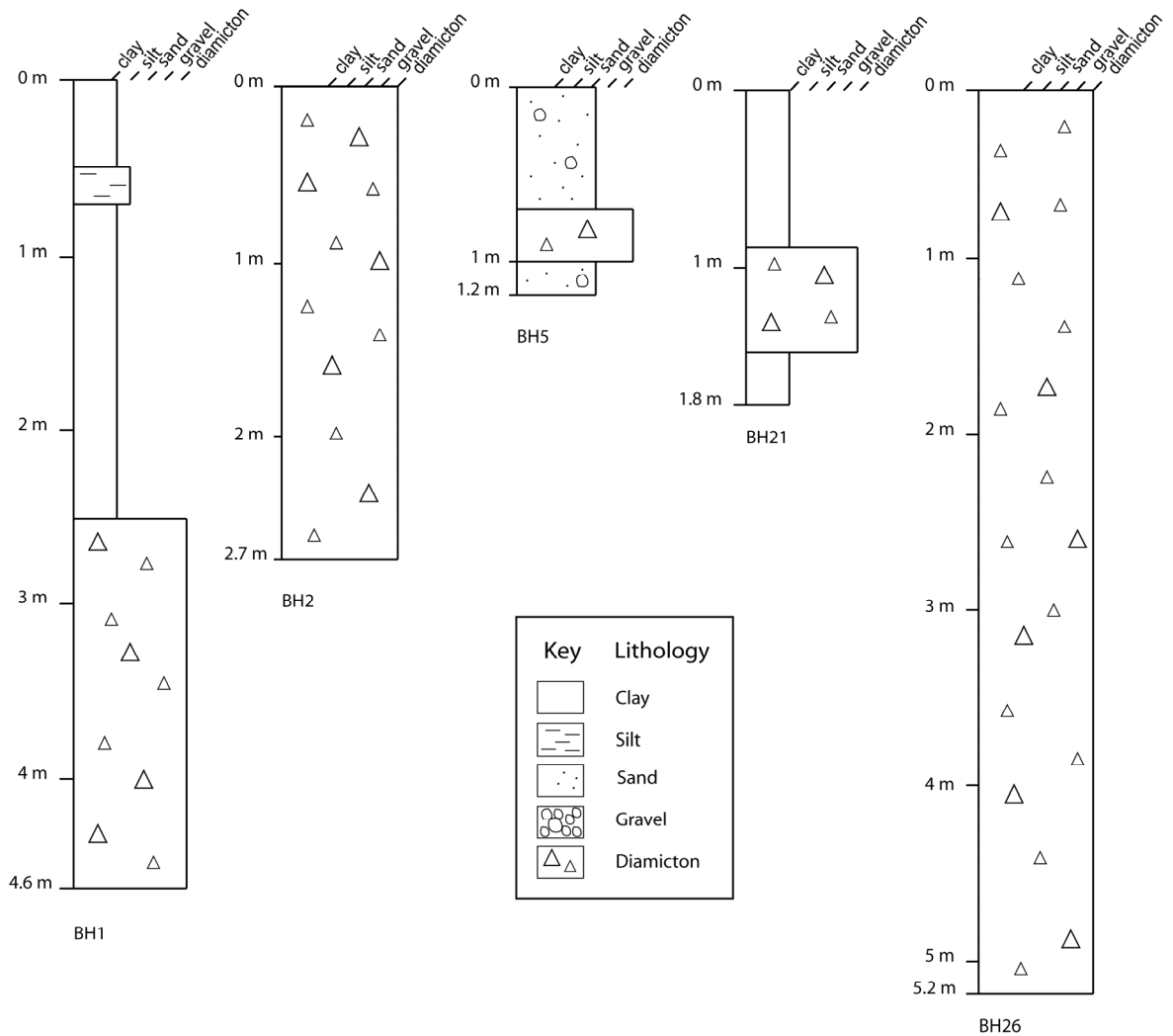


Figure 7.12: Borehole data (BGS) from the region SW of Wigton (see Fig. 7.1) showing disparate evidence for an upper clay or sand lithofacies interpreted to relate to the formation of a glacial lake.

The esker deposits at both Thursby and Sowerby Wood are also interpreted to have been deposited during the Scottish Re-advance. Although these features have been identified within the field (see Fig. 7.5), this interpretation has been derived from data collated from previous research (cf. Dixon *et al.*, 1926; Huddart, 1970, 1981). Clast lithological analysis reveals a Scottish provenance (Dixon *et al.*, 1926; Huddart, 1970), whilst the esker at Thursby is also characterised by a series of small hills (Dixon *et al.*, 1926; Huddart, 1970), not easily identified on the NEXTMap DEM. These hills are characterised by steep slopes along their western margin, giving way to lobate forms on their

eastern side (Huddart, 1970). They are interpreted as fan/delta deposits which constrain the position of the ice-margin during temporary still-stands, with water discharging eastwards into a subaqueous environment from subglacial tunnels (e.g. Banerjee & McDonald, 1975; Rust & Romanelli, 1975; Thomas, 1984; Gorrell & Shaw, 1991; Warren & Ashley, 1994).

### 7.5.2 *What was the maximum extent of the Scottish Re-advance?*

South of Carleton the tripartite sequence, comprising two tills separated by sand, gravel and laminated clay/silt lithofacies, has been interpreted to record the 'Blackhall Wood' Re-advance due to its stratigraphic position beneath lineations associated with flow out into the Irish Sea ice basin (Livingstone *et al.*, in press a). This tripartite sequence has previously been correlated with the Scottish Re-advance, and as such has resulted in an over-exaggeration of its maximum limit (Dixon *et al.*, 1926; Trotter 1929; Fig. 7.2a). Indeed, recent work carried out by Livingstone and co-workers (in press a) suggests that the region south of Carlisle is the sole preserve of the 'Blackhall Wood' till sequence.

The Wiza Beck meltwater channels are interpreted to have formed during the retreat stage of the 'Blackhall Wood' LT5 flow phase (e.g. Huddart, 1970), when ice was both downwasting and shrinking eastwards back into the Vale of Eden. Therefore, these meltwater channels do not constrain the southerly limits of the Scottish re-advance, as has been previously postulated (cf. Dixon *et al.*, 1926; Trotter, 1929). This interpretation is based on their complex morphology, characterised by sharp bends and undulatory long-profiles. Their morphology indicates either a subglacial origin (Sisson, 1960, 1961; Sugden *et al.*, 1991) or two-stage development, with the W-E orientated, parallel channel sections characteristic of an ice sub-marginal genesis (Sissons, 1961; Dyke, 1993). An ice sub-marginal interpretation is supported by a number of small sand and gravel deposits (up to 70 m O. D.) at the western end of the perched meltwater channels around Wigton (Eastwood *et al.*, 1968), which probably relate to delta/fan deposits formed as water debouched into a subaerial or subaqueous environment. Similarly, the meltwater channels at Biglands are interpreted as sub-marginal channels demarcating the retreat of flow-phase LT5 from Scotland.

To the east of Carlisle, the largest glacial feature of potential Scottish Re-advance age is the Brampton kame belt (Fig. 7.1), situated in the lee of the Pennines and comprising glaciofluvial and glaciolacustrine deposits up to 25 m thick, overlying bedrock or diamicton (Jackson, 1979; Huddart, 1970, 1981; Livingstone *et al.*, in press c). This feature is likely to have formed during the retreat of the ice from the Tyne Gap, prior to, or during the Blackhall Wood Re-advance (Livingstone *et al.*, in press c). The absence of upper till or overprinted landforms on the kame belt limits the Scottish Re-advance to the west of the landform.

Stratigraphic evidence east of Carlisle, indicates the existence of a discontinuous and thin diamicton sheet (<1.0 m) overlying the Crosby Moor delta (Huddart, 1970), observed at Botcherby clay pit (Trotter, 1929; Carruthers, 1932) and Eden Bridge (Huddart, 1994), and exposed in river sections of the Irthing, King Water and Cam (Trotter & Hollingworth, 1932). This diamicton sheet caps a series of glaciolacustrine deposits, which become exposed at the surface at Lanercost, Boothby Bank and Great Easby (Fig. 7.1). The stratigraphic position of the diamicton is certainly consistent with a Scottish Re-advance genesis (Trotter & Hollingworth, 1932b), while the lateral extent and thickness of this lithofacies, which also includes till at Plumpe Farm and Carleton, corroborates our earlier interpretation of subglacial deposition as a thin and patchy till sheet.

The glaciolacustrine deposits (BGS boreholes; Trotter & Hollingworth, 1932; Huddart, 1970) and deltas underlying this diamicton belt indicates the presence of a major glacial lake ('Glacial-Lake Carlisle') impounded against ice to the west of Carlisle. It is envisaged that the lake formed against the reverse slope of the Tyne Gap, with the westerly retreating ice margin resulting in falling lake levels (60 – 43 m) as recorded by a series of deltas at elevations which lower progressively towards the west. Thus, it is likely that 'Glacial Lake Carlisle' formed before the Scottish Re-advance, during gradual retreat of either the 'Main Glaciation' (corresponding to the Last Glacial Maximum) or 'Blackhall Wood' Oscillation. This is supported by the sections at Plumpe Farm and Carleton, both of which suggest significant glaciofluvial/glaciolacustrine activity during a period preceding the Scottish Re-advance.

The geometry and long profile of Wampool meltwater channel is indicative of a proglacial meltwater feature (cf. Benn & Evans, 1998), whereby large volumes of water were discharged westwards. It could therefore be associated with either (or both) the rapid over-spilling of Lake Carlisle (Huddart, 1970) during eastwards retreat of the Blackhall Wood Oscillation (flow-phase LT5: Chapter 2). Alternatively it functioned as a major proglacial channel throughout the late stage lobate re-advance of Vale of Eden ice into the fringe of the Solway Lowlands (flow-phase LT6: Chapter 2) (see research question 3).

Overall, it is thought that the re-advance of Scottish ice into the Solway Lowlands was less extensive than in some previous reconstructions (e.g. Trotter, 1929). With the discovery of the Blackhall Wood Re-advance, much of the evidence pertaining to the southerly expansion of the Scottish Re-advance can be rejected; several of the morphological features which have previously been correlated to the Scottish Re-advance are probably recessional landforms of the 'Blackhall Wood' Re-advance. The lack of exposures east of Carlisle has made it difficult to verify the existence of a subglacial till extending east as far as Lanercost (e.g. Trotter, 1929), although key sites at Plumpe Farm and Carleton, records from previously exposed sections and borehole evidence support our inference of deposition as a thin, patchy subglacial till sheet.

### 7.5.3 *Was there was a concurrent re-advance of ice from the Lake District?*

A flowset of south-north orientated subglacial lineations trending out of the Vale of Eden into the fringe of the Solway Lowlands records a late stage ice flow which occurred after the Blackhall Wood Re-advance (flow-phase LT6: Chapter 2). Till thickening in the northern zone of this flowset (Livingstone *et al.*, in press a) delimits a possible end moraine, characteristic of ice-marginal settings (Haeberli, 1981; Alley, 1991; Boulton, 1996a,b). This lobate re-advance may have occurred concurrently with the Scottish Re-advance, although such a correlation cannot be conclusively proven, and may alternatively represent an internal re-adjustment of the Lake District ice dispersal centre triggered by changes in regional ice-sheet geometry.

### 7.5.4 *What are the glaciological implications of the Scottish Re-advance?*

It is generally recognised that till thickens towards the sub-marginal zone of the glacier (Haeberli, 1981), with theoretical models proposing 'conveyor-belt' transport of sediment by sliding, deformation and/or ploughing (see Tulaczyk *et al.*, 2001), whereby the accumulation zone is dominated by erosion or net loss of sediment, and the ice-margins by deposition along a thin marginal band (Alley, 1991, 1997; Boulton, 1996a, b). Marginal thickening has been further attributed to glacial fluctuations producing “waves” of till to the ice margin (Boulton, 1996b), while Evans & Ó Cofaigh (2003) demonstrate glaciotectionic thickening by the folding and stacking of pre-existing sediments. Freeze-on of subglacial till also provides an effective process for transporting and then stacking sediments, and has been recognised as an effective process along the margins of Icelandic glaciers (Evans & Twigg, 2002; Evans & Hiemstra, 2005). The thin (<2 m), patchy till sheet and conspicuous lack of marginal moraines associated with the Scottish re-advance in the Solway Lowlands clearly contradicts this model.

Till thickening towards the sub-marginal zone of the ice mass occurs if sufficient time is available for ice-marginal fluctuations to deliver repeated “waves” of sediment (Boulton, 1996b), or if the margin remains stable for sufficient time for the incremental stacking of till to occur (Evans *et al.*, 2006). Thus the Scottish Re-advance probably represents a very short-lived pulse in accord with explanations given at other margins characterised by thin till sheets (e.g. Scourse *et al.*, 1990; Wingfield, 1994; Piotrowski & Kraus, 1997; Evans & Ó Cofaigh, 2003; Hiemstra *et al.*, 2006). This is supported by the lack of strandlines or burial of LT5 landforms by lake sediment from the glacial-lake which suggests that it was a short-lived event.

Microstructures present with the stratified base of the capping till and across the gradational contact into the underlying glaciolacustrine sediments at Plumpe Farm indicate that the sediments were water-saturated (Phillips *et al.*, 2007). Likewise, the gradational contact between glaciolacustrine sediment and till at Carleton implies similar conditions. Saturated sediments at the ice-bed contact

would have effectively buffered the transition of stresses into the underlying glaciolacustrine sand, silt and clay, with the ice able to flow rapidly across the water-lubricated sediments (e.g. Piotrowski, *et al.*, 2001). This interpretation is endorsed by the numerous glaciofluvial landforms associated with the Scottish Re-advance capable of producing large volumes of meltwater. The distinction between our observations and the sharp-contacts observed by Piotrowski and co-authors is explained by the difference between basal-sliding across a thin film of water (Piotrowski *et al.*, 2001) and the concentration of deformation within a water-saturated zone of sediment (e.g. Phillips *et al.*, 2007). Another controlling factor on the flow of ice across the Solway Lowlands could have been the influence of proglacial lakes within the region, with large volumes of ice-dammed meltwater capable of lowering basal shear stresses and triggering rapid ice motion by calving of the ice front (Stokes & Clark, 2004). A further compounding factor may have been that the ice sheet was debris poor.

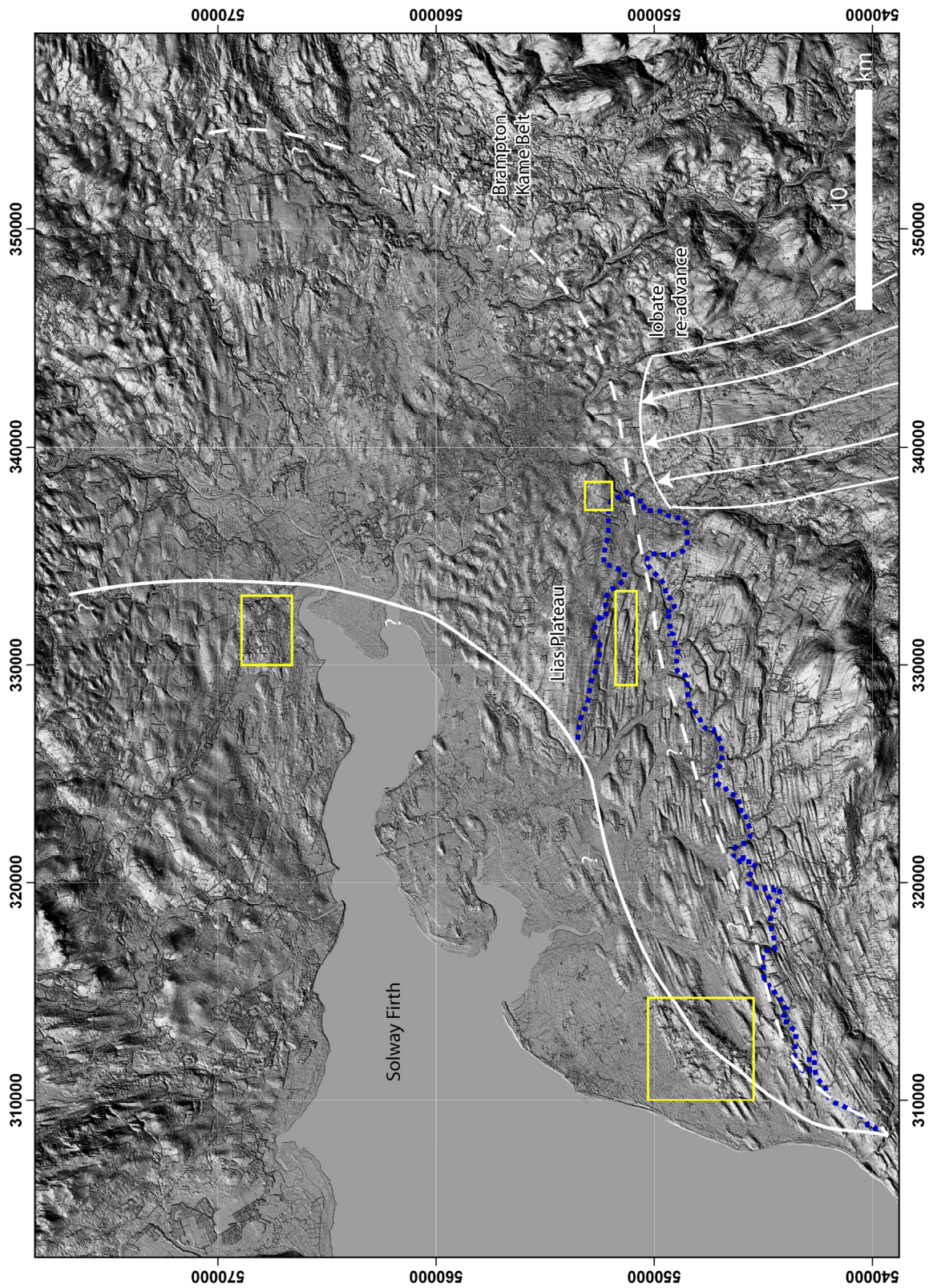
Given that the Scottish Re-advance deposited a thin-patchy till sheet and showed limited evidence for subglacial or ice-marginal landform generation, deformation and remoulding of underlying sediments it is therefore no surprise that the re-advance was a short-lived pulse characterised by movement over water-saturated sediment and into proglacial lakes.

## 7.6 Conceptual model of the Scottish Re-advance and its wider implications

The evidence presented in this paper has facilitated a refinement of the concept of a Scottish Re-advance into the Solway Lowlands (Fig. 7.13). Three major lines of evidence can be identified to summarise the advances made by this research.

Firstly, the Holme St. Cuthbert landform-sediment assemblage marks the most recognisable stillstand of Scottish Re-advance ice within the Solway Lowlands (Fig. 7.13). The dead ice morphology and size of the deltaic sequence indicate ice-marginal deposition during either a stillstand of significant duration or a phase of rapid deposition. A number of other features identified along the coast seemingly correlate with this depositional event, including the glacial outwash deltas and eskers at Gretna (cf. Stone *et al.*, in prep) at 30 m O. D., the St. Bees moraine complex (cf. Huddart, 1994; Merritt & Auton, 2000) and the proglacial sandur deposits at Harrington (cf. Huddart, 1994). Thus it is envisaged that the main stillstand of the Scottish Re-advance impinged on the Cumbrian coast, blocking drainage and subsequently leading to lake formation within the Wigton region (Fig. 7.13). Ice almost certainly advanced up to the Lias Plateau thus preventing water draining northwards into the Gretna region (Fig. 7.13). The foresets at Holme St. Cuthbert indicate a lake level of approximately 49 m O. D. (Fig. 7.13).





< Figure 7.13: Reconstruction of Scottish Re-advance into the Solway Lowlands. The yellow boxes refer to glacial landforms (glacial outwash delta and eskers at Gretna, Sowerby Wood and Thursby eskers and Holme St Cuthbert deltaic sequence) which can be confidently correlated with the Scottish re-advance. The white line demarcates the major stillstand position (see text), while the white dotted line refers to the rapid, transient lobe of thin ice, which spread out across the Solway Basin. The blue-dotted line extent of the lake associated with the Holme St. Cuthbert delta (delineated by the height of the foresets). The white arrows refer to the lobate advance of Vale of Eden ice into the Solway Lowlands.

Secondly, stratigraphic evidence presented in this paper demonstrates that the re-advance was a short-lived pulse which flowed over water-saturated sediments and into proglacial lakes, depositing a thin, patchy till sheet. This pulse reached as far east as the rivers Irthing and Cam (Dixon *et al.*, 1926; Trotter *et al.*, 1929; Trotter & Hollingworth, 1932; Huddart, 1970, 1994) and almost certainly extended beyond the Holme St. Cuthbert stillstand as far SE as the Thursby and Sowerby Wood eskers during an earlier ice-growth phase (Fig. 7.13). The short-lived nature of the re-advance which has resulted in “feather” edges precludes firm predictions of the maximum extent, although the identification of the Blackhall Wood Re-advance till member indicates that the southern-most limits suggested by Trotter (1929) were over-exaggerated (Fig. 7.13).

The cross-cutting patterns in the Solway Lowlands demonstrate that the 'Blackhall Wood' Re-advance, which has been tentatively linked to the Gosforth Oscillation (Merritt & Auton, 2000), occurred during an earlier stage of deglaciation (Livingstone *et al.*, in press a). Further cross-cutting of the Blackhall Wood flow-set by lineations splaying out into the Solway Lowlands from the Vale of Eden indicate that Lake District ice also underwent a subsequent re-advance, although without direct correlation it is hard to know whether this formed part of a regional re-advance signal or an internal re-adjustment. Further correlation of the NE sector of the Irish Sea Basin with the Clogher Head and Killard Point Stadials in NE Ireland (McCabe *et al.*, 2007) is difficult. Poor dating controls hinder attempts to assign known fluctuations to climatic events (e.g. the Killard Point Stadial and Heinrich event 1: McCabe *et al.*, 1998), while work in SW Scotland indicates that pulsed recession/surging in response to internal forcing, extensive marginal ponding and deepening marine water should not be dismissed as a possible mechanism for re-advances in the NE sector of the Irish Sea Basin (e.g. Salt and Evans, 2004; Stokes & Clark, 2004; Evans *et al.*, 2009). Although the Scottish Re-advance can be traced down the west Cumbrian coast (cf. Huddart, 1994; Merritt & Auton, 2000), evidence presented in this paper implies that it was a relatively minor and short-lived fluctuation with little erosive power. Indeed most of the evidence for the Scottish Re-advance is limited to glaciolacustrine/glaciofluvial deposits (such as Holme St. Cuthbert sediment-landform assemblage and Thursby and Sowerby Wood eskers) indicative of wide-spread lake development, dammed between ice in the Celtic Basin and upland regions inland (Fig. 7.13). These characteristics displayed by the Scottish Re-advance are more akin to a short-lived internally controlled fluctuation. Conversely, the Killard Point stadial, which has previously been correlated

with the Scottish Re-advance (McCabe *et al.*, 1998; McCabe & Clark, 1998), was thought to be more extensive. This is typified by the coeval advance of ice onto the Isle of Man (Bride moraine) and the drumlinisation and formation of rogen moraine throughout NE Ireland (McCabe *et al.*, 1998).

## 7.7 Conclusions

The complex, palimpsest landscape of the Solway Lowlands had hindered attempts to reconstruct and interpret individual ice-flow phases. For the Scottish Re-advance, a paucity of well-constrained landforms, sediments and dates has inevitably led to debate over its existence, limits and dynamic behaviour. This paper has combined both glacial geomorphic and sedimentological information to critically review the evidence for and against a late stage Scottish Re-advance into the Solway Lowlands, and to produce a refined reconstruction of the event (Fig. 7.13). Several key points can be drawn from this study:

1. Re-advance of Scottish ice into the Solway Lowlands did occur at a stage subsequent to the Main Glaciation and Blackhall Wood Re-advance. This is supported by disparate stratigraphic evidence of an upper till sheet, a series of eskers along the edge of the Lias Plateau and the Holme St Cuthbert deltaic sequence.
2. A major still-stand recorded by the Holme St. Cuthbert delta sequence and traced along the Cumbrian coast marks the most obvious limit of the Scottish re-advance. At this time a large ice-contact lake formed in the vicinity of Wigton. Subdued eskers and evidence of a thin, patchy upper till sheet points towards a transient ice advance further inland (Fig. 7.13) that is not thought to relate to the Holme St. Cuthbert still-stand. A lobate re-advance of ice down the Vale of Eden may have been concurrent with the Scottish Re-advance.
3. The thin till sheet observed at the margins of the Scottish Re-advance is in direct contrast to conceptual models and observed trends of sub-marginal till thickening (cf. Boulton 1996a, b). This thin till sheet is attributed to a short-lived re-advance pulse which probably moved rapidly across water-saturated sediment and into proglacial lakes dammed up against higher ground to the east of the ice. The short-lived nature of the re-advance coupled with the buffering effects provided by water-saturated sediment at the ice-bed interface meant the re-advance exerted little depositional, erosional or deformational influence.
4. Other glacial features that have in the past been assigned to the Scottish Re-advance, including the Wampool and Wiza Beck meltwater channels, an upper till south of Carlisle and the development of Glacial-Lake Carlisle are re-interpreted here as belonging to earlier to flow phases.

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## Chapter 8

# **The application of geochemistry and particle size analysis to the investigation of glacial stratigraphy in the central sector of the last British-Irish Ice Sheet**

### **Abstract**

The use of geochemical investigations to constrain ice-flow histories has been successfully demonstrated in Fennoscandia and North America, yet only sporadically utilised within the U.K. This chapter applies geochemical and particle size analysis, using exploratory statistical techniques to the investigation of glacial stratigraphy in the glacially and geologically complex Solway Lowlands and Tyne Gap region of NW England. The techniques used in this chapter successfully distinguish and therefore verify stratigraphic units identified independently using geomorphological and sedimentological methods, both in vertical sections and across the region. Five major geochemical assemblages are recognised from this analysis, which discriminate between Scottish, Lake District-Vale of Eden, mixed Lake District-Vale of Eden and Scottish, mixed Lake District-Vale of Eden and Tyne Gap and local provenances. These assemblages have allowed ice-flow pathways to be identified, thus confirming the existence of Scottish ice along the west Cumbrian coast either before or during the Blackhall Wood re-advance; while the identification of an upper till with a Scottish provenance collaborates the re-advance of Scottish ice to the SE edge of Carlisle. The geochemical similarities between diamicton lithofacies of the tripartite stratigraphy in the northern sector of the Vale of Eden indicates similar ice-flow pathways sourced from the Lake District-Vale of Eden. This study also demonstrates the strong influence that local bedrock had on the provenance signal, thus indicating that the cannibalisation of underlying glacial sediments and rafting-up of underlying bedrock were significant processes.

### **8.1 Introduction**

Geochemical exploration has been used extensively in both Fennoscandia and Canada since the 1960s as a tool for tracing minerals back to their source ore (Shilts, 1993). A bi-product of this industry-driven technique has been the identification of glacial dispersal trains, which in turn has led to a greater understanding of ice-flow directions within ancient ice sheets (e.g. Saarnisto, 1990; Shilts *et al.*, 1979; Dyke *et al.*, 1982; Shilts, 1993). Research has mainly been associated with the lateral extent of glacial dispersal trains, although some geochemical investigations have also been carried out on vertical sections in order to differentiate between stratigraphic units (e.g. May & Dreimanis, 1976; Steele *et al.*, 1989). Glacial research carried out in the U.K. has been slow to

apply geochemical investigations. However, the limited studies which have been carried out have demonstrated the potential of this technique for reconstructing ice-flow histories (Burek & Cubitt, 1979, 1991; Walden *et al.*, 1987, 1992, 1995; Richards, 2002; Davies *et al.*, 2009).

The Late Devensian (Marine Isotope Stage 2) glacial history of the Solway Lowlands and Tyne Gap (Fig. 8.1) is complex, reflecting multi-phase ice-flow fed by competing ice-dispersal centres over the Lake District and Southern Uplands (Huddart & Glasser, 2002; Livingstone *et al.*, 2008, in press a; Evans *et al.*, 2009; Stone *et al.*, in press). Together with a paucity of exposures this has made it difficult to correlate stratigraphies between field sites. This paper therefore aims to apply geochemical techniques and statistical analysis to a geologically and glacially complex region of the British and Irish Ice Sheet (BIIS) in order to reconcile stratigraphic complexities, both in vertical sections and over large areas. The new quantitative data derived from this work will allow us to critically evaluate ice-flow histories, debris provenance patterns and stratigraphic correlations within the Solway Lowlands and Tyne Gap.

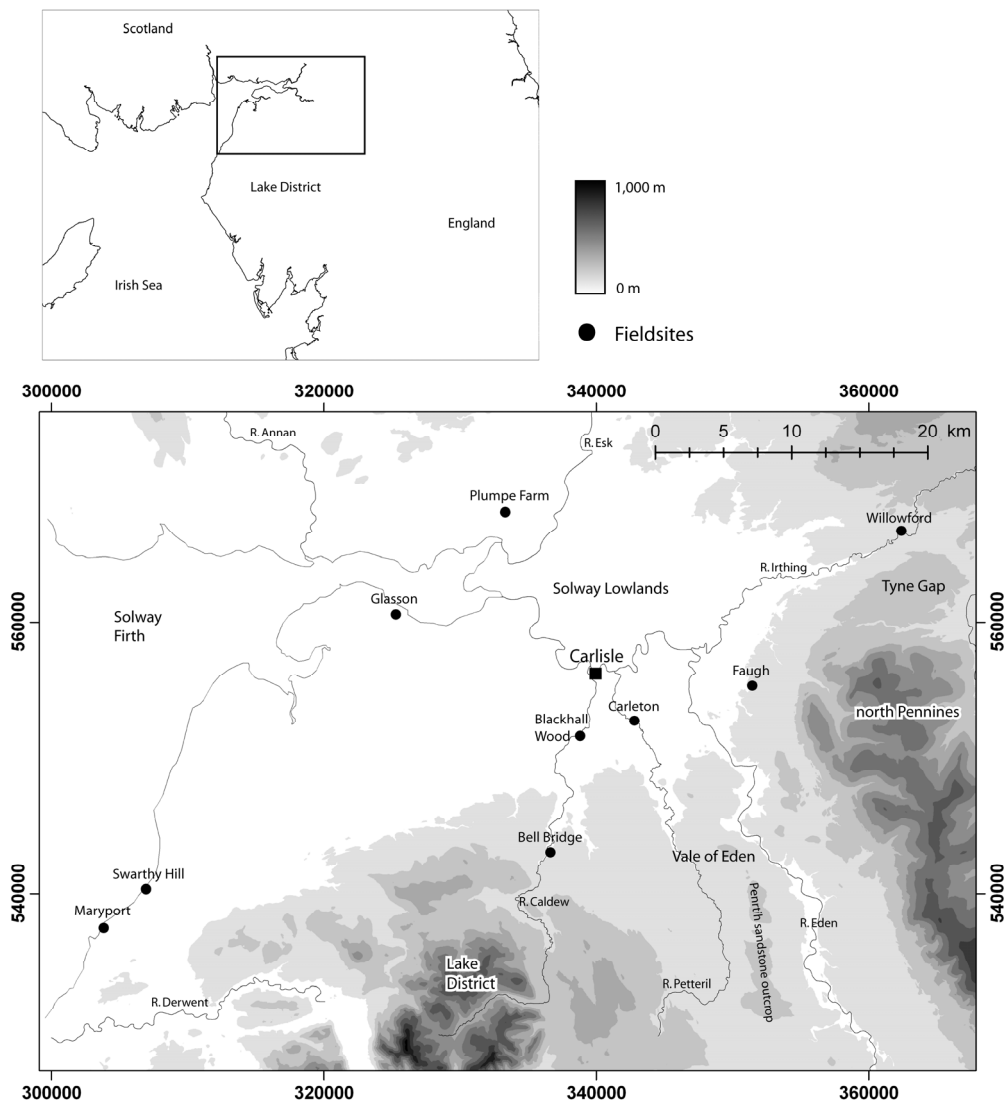


Figure 8.1: Location Map, including field sites.

## 8.2 Physiographic setting

### 8.2.1 *Bedrock geology*

The varied geology exhibited within north-western England and south-western Scotland has resulted in a geomorphologically complex landscape (Fig. 8.2). The north Pennines (Alston Block) is composed of Carboniferous rocks, including shale, limestone, sandstone and Millstone grit, which dip gently towards the east (King, 1976). It has well-defined and faulted western (Pennine fault) and northern (Stublick fault) margins. Uplift of the Alston Block during the Hercynian earth movements brought the underlying Lower Palaeozoic rocks to the surface at Cross Fell, creating a geological inlier, while this process also resulted in the formation of the faulted Pennine escarpment (King, 1976). The Lake District is a dome of Lower Palaeozoic rocks, with three bands, which from north to south comprise Skiddaw slates of Ordovician age, the Borrowdale Volcanic Series, and Silurian shales, grits, mudstones and siltstones which form rather more subdued relief (King, 1976). A number of igneous intrusions including Shap granite, Ennerdale grannophyre, Eskdale granite, Threlkeld syenite and Carrock Fell gabbro, also form a series of outcrops throughout the Lake District (Trotter, 1929). The outer edges of the dome are composed of Carboniferous limestone which dips gently eastwards and northwards beneath the Permo-Triassic sandstone of the Vale of Eden (Trotter, 1929). The Southern Uplands primarily consists of Ordovician and Silurian grits, slates and greywacke, with some bands of chert and intrusions of granite (Charlesworth, 1926). The major lowland regions of north-west England include the Solway Lowlands, Vale of Eden and Tyne Gap (Fig. 8.1). Much of the Vale of Eden and Solway Lowlands consist of Permian and Triassic sandstones (Fig. 8.2) (King, 1976). The lowest division of this is the Penrith sandstone which only outcrops in the Eden Valley, forming a prominent 24 km long ridge (Fig. 8.1). The Tyne Gap is underlain by Carboniferous sedimentary rocks (Fig. 8.2) consisting of limestone, shale, sandstone and coal that dip southwards towards the edge of the Alston Block (King, 1976). More resistant rocks, including the Great Whin Sill, result in strong W-E orientated cuestas.

### 8.2.2 *Quaternary stratigraphy*

The lowland regions of northwest England are characterised by thick successions of glacial drift related to the Late Devensian glaciation (cf. Huddart & Glasser, 2002; Stone *et al.*, in press). Early research identified three main stages of ice-flow (Fig. 8.3): an early ‘Scottish Advance’; a later ‘Main Glaciation’; and then a re-advance phase (Trotter & Hollingworth, 1932).

The early ‘Scottish Advance’ was characterised by movement of Scottish ice up the Vale of Eden and across the Stainmore Gap, with Lake District ice also being forced eastwards across this mountain pass (Fig. 8.3) (Trotter, 1929; Hollingworth, 1931; Huddart, 1970; Catt, 2007). Evidence for this advance is limited to a lowermost, disparately exposed division of reddish-brown till (Gillcambon Till Formation) containing Scottish clasts and the widespread dispersion of Scottish

erratics throughout the region (Table 8.1) (Trotter, 1929; Hollingworth, 1931). Recent research (Evans *et al.*, 2009; Livingstone *et al.*, in press a) questions the glaciological plausibility of this advance and instead proposes that local ice-flows may have re-distributed Scottish erratics transported to NW England during earlier glacial events.

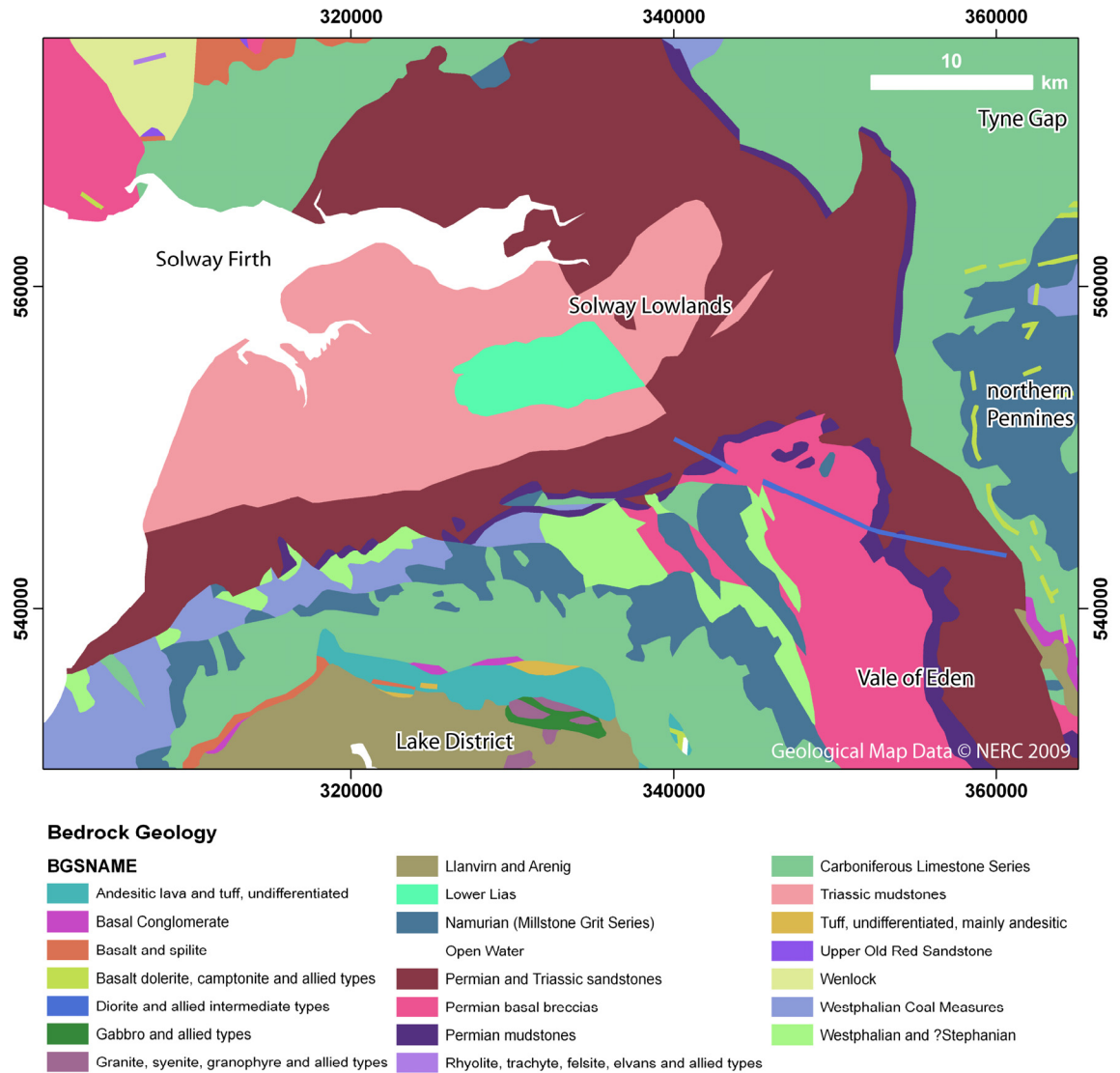


Figure 8.2: 1:625k bedrock geology map of the Solway Lowlands and Tyne Gap.

Following glacial re-organisation (Middle Sands), which may have resulted in the partial deglaciation of the Vale of Eden (Table 8.1) (cf. Hollingworth, 1931: laminated clays at Langwathby), the ‘Main Glaciation’ was characterized by ice-flow from west to east across the country, through major mountain passes in the northern Pennines (the Stainmore and Tyne Gaps) (Trotter, 1929; Hollingworth, 1931). This corresponds to flow-phases ES1 and ST1-3 of the Stainmore Gap (see Chapter 2, Fig. 2.5; Table 8.1). Flow through the Tyne Gap occurred as a topographic ice-stream (Fig. 8.3), sourced from Lake District and Scottish ice converging on the Solway Lowlands (Livingstone *et al.*, in press b) and relates to flow-phases LT1-3, outlined in Chapter 2 (Fig. 2.5). Stratigraphic evidence for the ‘Main Glaciation’ is recorded by an extensive red till in the Vale of Eden (Greystoke Till Formation; Table 8.1) (Trotter, 1929; Hollingworth,

1931). Till from the 'Main Glaciation' is also prevalent within the Tyne Gap, although it is characterised by a gradual change in colouration (from red to grey) towards the east and north, related to the influence of Scottish Southern Uplands rocks and local Carboniferous limestone (Livingstone *et al.*, in press b). In the northern sector of the Vale of Eden this stratigraphy is complicated by a tripartite sequence of glacial sediments, characterised by an upper till (LF5 of Blackhall Wood, Chapter 6) and lower till (LF 1 and LF2 of Blackhall Wood, Chapter 6) separated by sand, silt and clay deposits. The lower tills have been attributed to the 'Main Glaciation', and the upper till to the Blackhall Wood Oscillation during which ice re-advanced into the Irish Sea Basin (flow-phase LT5: Chapter 2) following the development of ice-free conditions within the Solway Lowlands (Table 8.1) (Livingstone *et al.*, in press a). Finally, a Scottish re-advance into the Solway Lowlands has been identified (Fig. 8.3) on the basis of a thin upper till (Gretna Till Formation) (overlying a series of clays, sands and gravels, deposited during the retreat of the 'Blackhall Wood' re-advance) and a series of eskers and ice-contact deltas in the north of the region (cf. Huddart & Glasser, 2002; Stone *et al.*, in press) (see also Table 8.1 and Chapter 7 above). Flow-phases SF1 and LT6 have been assigned to the Scottish re-advance (Chapter 2, Fig. 2.5; Table 8.1).

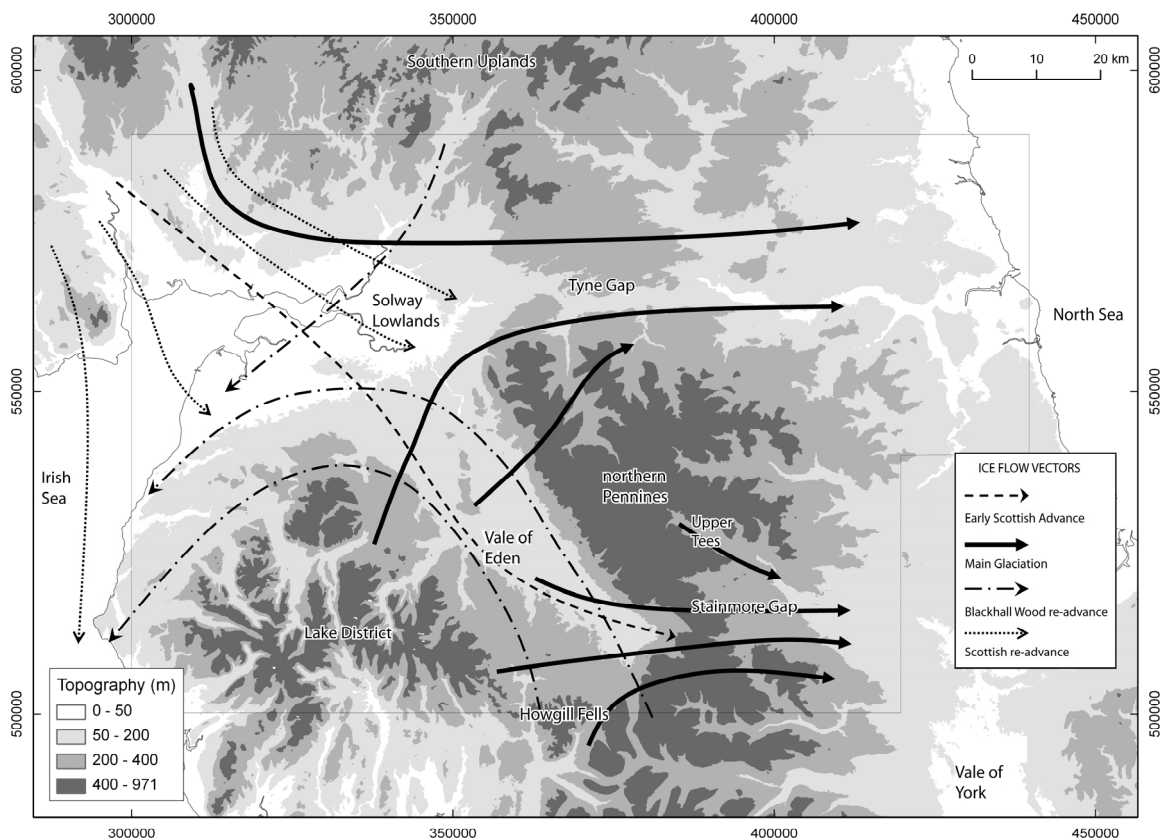
Event	BGS Lithostratigraphic Formations		Geomorphology: flow-sets (Chapter 2)
	Cumbria	Dumfries-shire	
Scottish Re-advance	Gretna Till Formation		SF1; LT6
Deglaciation (formation of Glacial-Lake Carlisle)	Great Easby Clay Formation	Plumpe Sand and Gravel Formation	
Blackhall Wood Re-advance	Greystoke Till Formation	Chapelknowe Till Formation	LT5
Deglaciation			
Main Glaciation			LT1-3; ST1-3
Middle Sands			
Early Scottish Advance	Gillcambon Till Formation		

Table 8.1: Simple Event Stratigraphy for the Solway Lowlands, showing BGS Lithostratigraphic Formations and glacial geomorphological flow-sets (from Livingstone *et al.*, 2008).

The stratigraphies of the nine field sites used in this study are illustrated in Fig. 8.4 and described in turn below. Lithofacies codes, illustrated in Fig. 8.4 and utilised throughout the Chapter, are based upon those of Evans & Benn (2004). Detailed descriptions and interpretations of these sites are also found in Livingstone *et al.* (in press a, b, c) and see also Chapters 4-7. Willowford comprises glaciofluvial deposits (Gm, Gms, Bms), overlain by glaciolacustrine deposits (Sd), debris-flows (Dms) and then intercalated reddish and brownish-black subglacial traction tills of the 'Main Glaciation' (Dmm) (Livingstone *et al.*, in press b). Rafts of crushed bedrock have been recorded within this upper till unit (sample W5). Blackhall Wood is the type location for the 'Blackhall Wood' re-advance and is characterised by two lower subglacial traction tills (Dmm) related to the flow of ice northwards down the Vale of Eden and then eastwards into the Tyne Gap (flow-phases LT1-3: Chapter 2). The upper till grades into debris-flows (Dms, Sm) that are analogous to sedimentation in a proglacial environment (Livingstone *et al.*, in press a). The debris-flows are



overlain stratigraphically by varved sediments (Flv), deposited in a proglacial lake, which in turn are capped by a thick subglacial traction till (Dmm) corresponding to the 'Blackhall Wood' re-advance (flow-phase LT5: Chapter 2; Livingstone *et al.*, in press a). The stratigraphy of Faugh sand pit has been interpreted as the product of debris-flow deposition (Dmm, Sm, Sd) into an ice-walled lake (Sp), during stagnation and *in situ* down-wasting of the ice margin (Livingstone *et al.* in press c). The exposure at Glasson displays a massive, 8 m thick till (Dmm) exposed in the 'stoss' end of a ENE-WSW orientated drumlin assigned to flow-phase LT5 (Chapter 2; Livingstone, unpublished work). Sediments exposed at Plumpe Farm comprise an upper subglacial traction till (Dmm), associated with the Scottish re-advance, overlying glaciectonised glaciofluvial/lacustrine deposits (Sm, Sd, Fm) (Phillips *et al.*, 2008). The lithofacies at Maryport are consistent with an origin as a glaciectonite (Dmm, Dms), derived from underlying glacio-fluvial/lacustrine sediments (Sd, Gms) and moulded into a drumlinoid form (Livingstone *et al.*, in press a). Bell Bridge consists of a lower glaciectonite (Dmm), which has rafted-up and crushed the soft underlying Westphalian and Stephanian mudstone, capped by an upper diamicton (Dmm). The exposure at Carleton displays glaciofluvial and deltaic sediments (Gruc, Gm, Gms, Fd, Sd). These are capped by a thin, discontinuous till (Dmm) thought to relate to the Scottish re-advance phase (see Chapter 7 above). Exposures through the drumlin at Swarthy Hill suggests that it formed by glaciectonisation of pre-existing glacio-fluvial/lacustrine sediments (Gms, Sl, Fl, Sd(w)) and debris flow deposits during ice-sheet over-riding (Dmm) (Livingstone *et al.*, in press a).



< Figure 8.3: Regional ice flow dynamics in the central sector of the BIIS during the Main Late Devensian (based on Trotter, 1929; Livingstone *et al.*, 2008, in press a). Box represents extent of geomorphological mapping by Livingstone *et al.*, 2008).

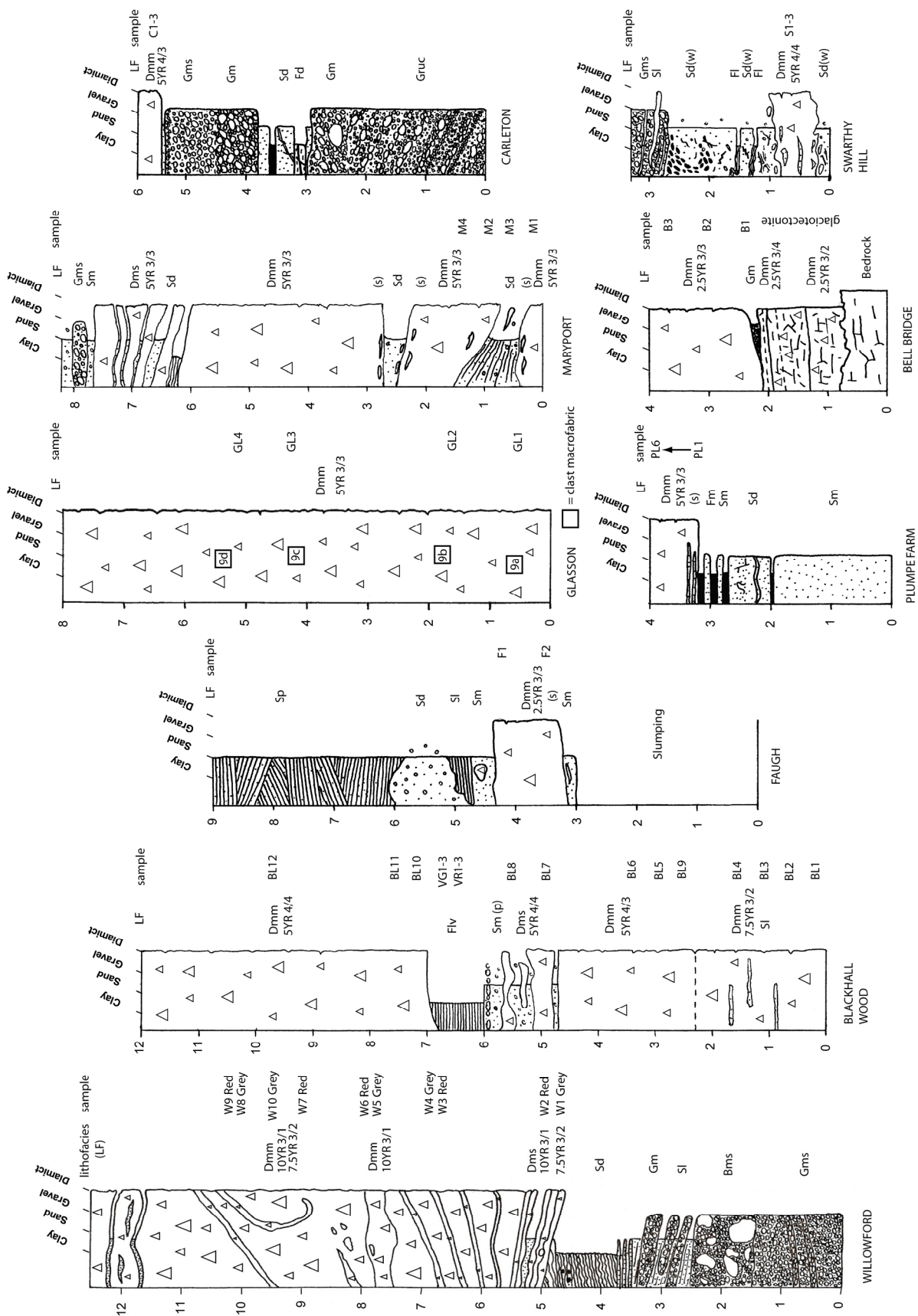
### 8.3 Methods

#### 8.3.1 Sample preparation and analysis

Geochemical and particle size analysis was carried out on all diamicton units and varved sediments (cf. Livingstone *et al.*, in press a) identified within the Solway Lowlands and Tyne Gap. Fig. 8.4 illustrates the sampling strategy and labelling procedure for geochemical and particle size analysis at each field site. A total of fifty-three samples were taken from nine sections, of which two were characterised by multiple diamicton lithofacies (see Figs. 8.1 and 8.4). The two sections with multiple diamicton lithofacies, at Willowford and Blackhall Wood (Fig. 8.4), were sampled at 0.5 m vertical intervals, while the smaller exposures required more detailed sampling in order to obtain a representative data set. Geochemical investigations within the U.K. remain a relative means of differentiating lithofacies because of the sparse network of samples available for the region and also because geochemical fingerprinting of bedrock sources has yet to be systematically carried out. Therefore, although geochemical signatures can be used to group samples, inferences about bedrock source has to be made in association with glacial flow-sets, macrofabric data and erratic evidence.

Determination of the geochemical content was carried out using inductively coupled plasma mass spectrometry (ICP-MS), whilst particle size analysis on the sub 2 mm fraction of the diamicton matrix was conducted using a laser granulometer. Samples for geochemical analysis were prepared as follows: samples were freeze dried for 24 hours before being crushed and homogenised in a ball mill. Hydrogen peroxide was added to remove any biological material. The samples were then extracted using 9 ml of HNO<sub>3</sub>, 3 ml of Hydrochloric acid and 2 ml of Hydrofluoric acid, with samples subsequently placed in a MARS pressurised microwave extraction system for 10 minutes at 180°C. Following this, the samples were filtered, and two separate runs were made with the ICP-MS, one for low abundance elements and one for high abundance elements. A suite of internal standards were used for each run. Sampling and preparation of samples for particle size analysis used standard procedures as outlined by Gale & Hoare (1991). Output from the particle size analysis is displayed as ternary plots which graphically demonstrate the percentage of clay, silt and sand within each sample.

## CHAPTER 8: GEOCHEMICAL AND PARTICLE SIZE ANALYSIS



< Figure 8.4: Composite stratigraphic logs of all the field sites, plus locations and names of samples taken for geochemical and particle size analysis.

### 8.3.2 *Statistical analysis*

Principle component analysis (PCA) has been widely applied in glacial stratigraphic investigations (Gibbard, 1986; Cheshire, 1986; Kovach, 1995; Richards, 1998, 2002; Davies *et al.*, 2009). It aims to simplify large multivariate data sets by reducing the information into a series of components, each of which reveals some form of underlying structure (Davis, 1986). All components which have an eigenvalue (measure of the relative importance of a component) of greater than 1.0 were retained (Davis, 1986). These components were plotted as scattergrams (score plots) with the output revealing how variables or samples relate to each other in bivariate space, and the relative importance of these relationships (e.g. Richards, 2002).

Hierarchical cluster analysis (HCA) is used to group multivariate observations into a number of meaningful assemblages, which display similar data structures (Kovach, 1995; Templ *et al.*, 2008). The output of HCA is the dendrogram, which is used to group either the samples or variables by visual observation. This statistical approach has been used as an exploratory tool in this study, with various methods (Ward, complete linkage and average linkage) used to assess the geochemical similarity of samples and variables. The Euclidean distance measure (size of dissimilarity between clusters) was used within this study as it is known to produce meaningful results (cf. Templ *et al.*, 2008).

Elements Tl, Th and Nd which have values below the detection limit were omitted from the analysis, while data with a heavy skew ( $> \pm 2.0$ ) were log-transformed to a more symmetrical distribution (cf. Templ *et al.*, 2008). As the variables show a significant difference in the amount of variability (major, minor and trace elements), which can heavily influence the weighting of the higher magnitude elements, the data was standardised (Templ *et al.*, 2008). The most common method for standardising the data, and the one used in this study, is the z-transformation (gives new  $z_i$  values), in which the raw data is subtracted from the mean and then divided by the standard deviation (Templ *et al.*, 2008).

## 8.4 **Results**

### 8.4.1 *Geochemical analysis*

PCA and HCA were carried out on the entire transformed and standardised dataset. Table 8.2 presents the 6 components retained from PCA. The first two components account for 53% and 15% of the variance respectively and thus are responsible for the majority of the variance exhibited by

the dataset. Components 3 - 6 are less important, accounting for between 3.6 and 5.6% of the explained variation each (Table 8.2). Component 1 is characterised by high positive loadings in Li, V, Co, Ga, Sn, U, Fe and Ti, whilst component 2 has high positive loadings in Ca, Ce, Mg, Sr and Y and large negative loadings in Al, P and Rb (Table 8.2 & Fig. 8.5). Component 3 has a very high loading in Ba and K (Table 8.2).

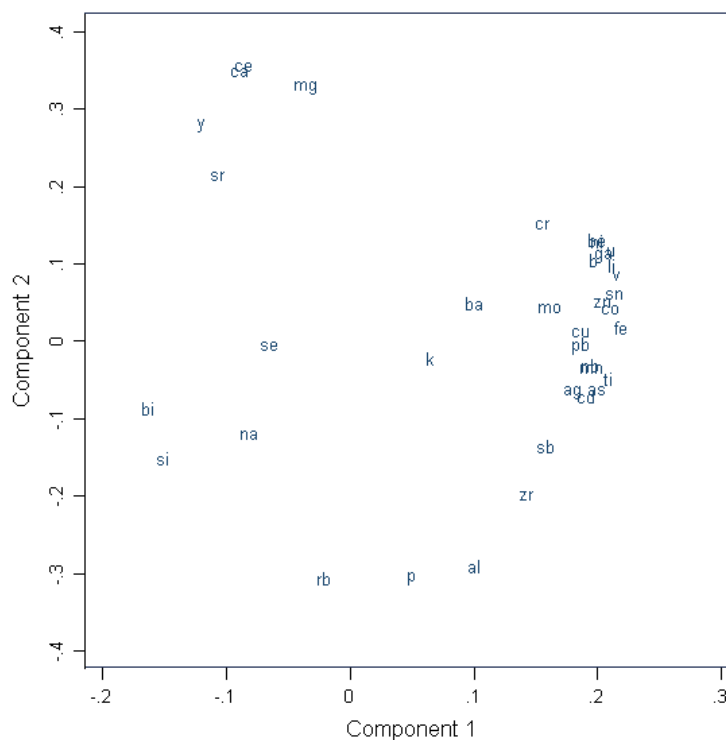


Figure 8.5: PCA: Plot of loading for the first two components.

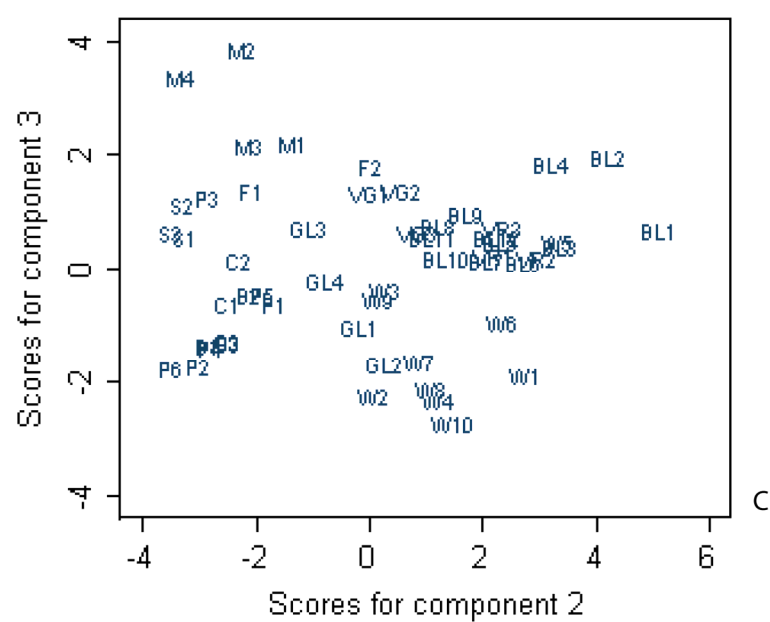
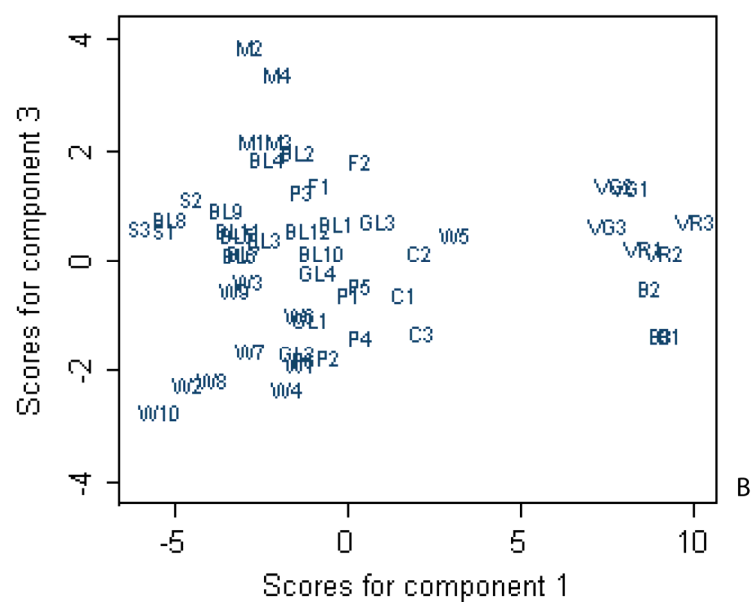
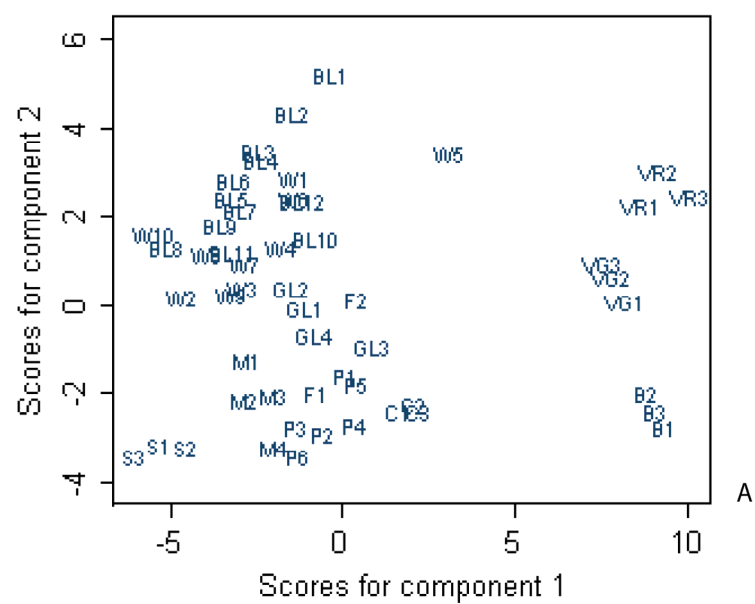
Score plots between components 1 and 3 reveal a number of groupings within the dataset (Fig. 8.6). The varves (samples VG1-3; VR1-3: Fig. 8.4) are easily distinguishable (Fig. 8.6a, b) with high positive loadings associated with component 1; indeed in Fig. 8.6a it is possible to separate out the signature of the summer and winter part of the couplets, most likely related to the hydrodynamic properties of constituent minerals (Briggs et al. 1962). The diamicton at Bell Bridge (samples B1-3: Fig. 8.4) also has a high positive loading associated with component 1, which makes it easily recognisable. The Swarthy Hill diamicton (samples S1-3: Fig. 8.4) is distinguishable in Fig 8.6a based on negative scores for both components 1 and 2, whilst the Maryport diamicton samples (M1-4: Fig. 8.4) (particularly M2 and M4) have high positive loadings in component 3 (Fig. 8.6b, c). There is very little difference between the samples from Carleton (C1-3: Fig. 8.4), Plumpe Farm (PL1-6: Fig. 8.4), Glasson (GL1-4: Fig. 8.4) and Faugh (F1-2: Fig. 8.4) which therefore means that they can be grouped with confidence as being of the same provenance (Fig. 8.6). The diamicton samples at Willowford (W1-4 & W6-10: Fig. 8.4) are very similar, although W5 has a much higher positive loading in component 1 (Fig. 8.6a, b). The samples from Blackhall Wood (BL1-12: Fig. 8.4) exhibit a positive relationship between components 1 and 2, with lower numbered samples

(e.g. BL1 and 4) associated with higher scores (Fig. 8.6a). In general, however, the Blackhall Wood and Willowford samples seem to occupy similar bivariate spaces (Fig. 8.6).

Variable	Component 1	Component 2	Component 3	Component 4	Component 5	Component 6
<i>Se</i>	-0.065	-0.004	0.194	<b>0.473</b>	0.049	-0.138
<i>Zr</i>	0.143	-0.197	<b>-0.201</b>	0.145	<b>0.308</b>	-0.005
<i>Rb</i>	-0.020	<b>-0.307</b>	<b>0.204</b>	-0.026	-0.011	<b>0.461</b>
<i>Sr</i>	-0.106	<b>0.215</b>	0.070	0.007	<b>0.210</b>	<b>0.457</b>
<i>Bi</i>	-0.163	-0.089	<b>-0.211</b>	0.124	-0.179	<b>0.238</b>
<i>Ba</i>	0.101	0.047	<b>0.481</b>	0.141	0.104	0.114
<i>Al</i>	0.101	<b>-0.291</b>	0.148	0.150	-0.107	0.173
<i>Ca</i>	-0.089	<b>0.349</b>	-0.006	-0.123	0.009	<b>0.211</b>
<i>K</i>	0.066	-0.024	<b>0.493</b>	0.059	<b>-0.230</b>	0.180
<i>Na</i>	-0.081	-0.120	<b>0.307</b>	<b>-0.425</b>	<b>0.244</b>	<b>-0.267</b>
<i>Si</i>	-0.151	-0.153	0.141	0.193	0.174	-0.087
<i>Li</i>	<b>0.212</b>	0.097	-0.041	-0.012	-0.011	0.105
<i>Be</i>	<b>0.200</b>	0.129	0.121	-0.012	-0.012	0.024
<i>B</i>	0.198	0.104	-0.013	<b>0.206</b>	0.002	0.183
<i>V</i>	<b>0.216</b>	0.087	-0.008	0.021	0.013	-0.057
<i>Cr</i>	0.156	0.152	-0.010	<b>0.309</b>	0.134	-0.044
<i>Mn</i>	0.195	-0.034	-0.024	-0.148	-0.079	0.036
<i>Co</i>	<b>0.211</b>	0.043	-0.040	-0.004	0.077	0.014
<i>Ni</i>	<b>0.200</b>	0.129	-0.033	0.129	-0.001	-0.058
<i>Cu</i>	0.187	0.012	-0.002	-0.074	-0.158	-0.113
<i>Zn</i>	<b>0.204</b>	0.050	-0.077	0.052	-0.031	-0.060
<i>As</i>	0.199	-0.062	0.073	-0.164	-0.145	0.025
<i>Mo</i>	0.162	0.044	-0.110	-0.194	0.077	<b>0.234</b>
<i>Ag</i>	0.180	-0.061	-0.018	<b>-0.260</b>	<b>0.292</b>	0.110
<i>Cd</i>	0.191	-0.072	-0.114	-0.057	<b>0.212</b>	0.036
<i>Sb</i>	0.158	-0.136	<b>0.224</b>	<b>-0.231</b>	-0.075	-0.137
<i>Ga</i>	<b>0.205</b>	0.115	0.181	0.023	-0.050	-0.006
<i>Sn</i>	<b>0.214</b>	0.061	0.052	-0.050	0.012	-0.047
<i>Y</i>	-0.119	<b>0.283</b>	0.065	-0.015	<b>0.334</b>	0.123
<i>Pb</i>	0.186	-0.005	-0.074	0.096	<b>-0.278</b>	0.066
<i>U</i>	<b>0.211</b>	0.118	0.037	-0.036	-0.049	0.003
<i>Ce</i>	-0.086	<b>0.356</b>	-0.041	-0.169	-0.070	0.175
<i>Nb</i>	0.195	-0.032	-0.102	0.189	<b>0.231</b>	0.016
<i>Fe</i>	<b>0.219</b>	0.017	-0.065	-0.035	-0.051	-0.008
<i>Mg</i>	-0.035	<b>0.331</b>	<b>0.223</b>	0.066	0.188	<b>-0.258</b>
<i>Ti</i>	<b>0.209</b>	-0.050	-0.037	0.054	0.108	-0.075
<i>P</i>	0.051	<b>-0.303</b>	-0.014	-0.017	<b>0.374</b>	0.144
<i>Eigenvalue</i>	14.116	5.510	2.082	1.719	1.481	1.341
<i>Explained variance (%)</i>	53	15	5.6	4.6	4	3.6
<i>Cumulative % of variance</i>	53	68	73.6	78.2	82.2	85.8

Table 8.2: Principle Components (bold numbers relate to the largest positive and negative loadings).





< Figure 8.6: PCA: Score plots for components 1-3: (a) component 1 vs component 2; (b) component 1 vs component 3; and (c) component 2 vs component 3.

HCA was carried out on both the variables and samples, with dendograms produced to display the output (Figs. 8.7 & 8.8). Fig. 8.7 presents clusters of elements which display similar behaviours in characterising clusters of samples. Ca, Ce, Mg, Sr and Y seem to be closely related (also suggested by PCA), and the samples identified from component 1 (PCA) are also clustered (Fig. 8.7). The cluster composed of Rb, P, Al, Ba and K (Fig. 8.7) is characterised by large negative loadings of component 2 and large positive loadings of component 3 (PCA).

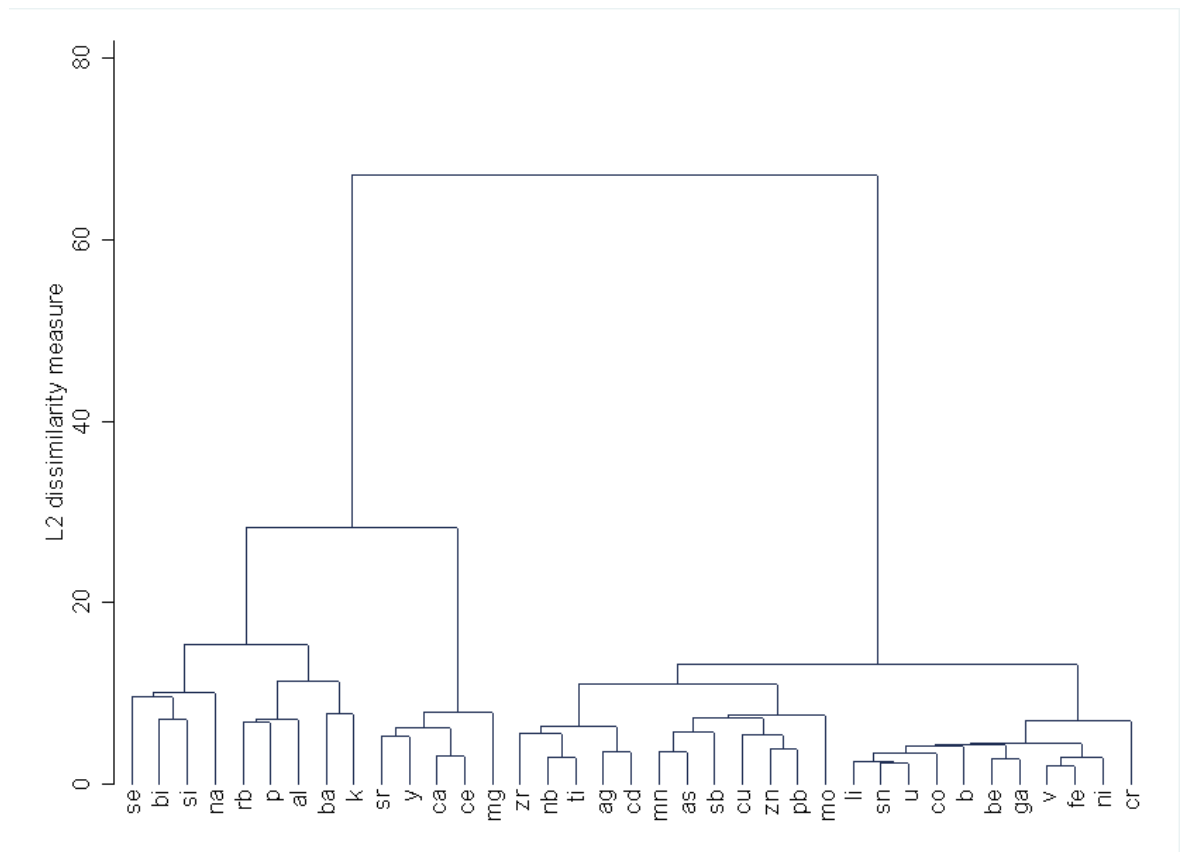
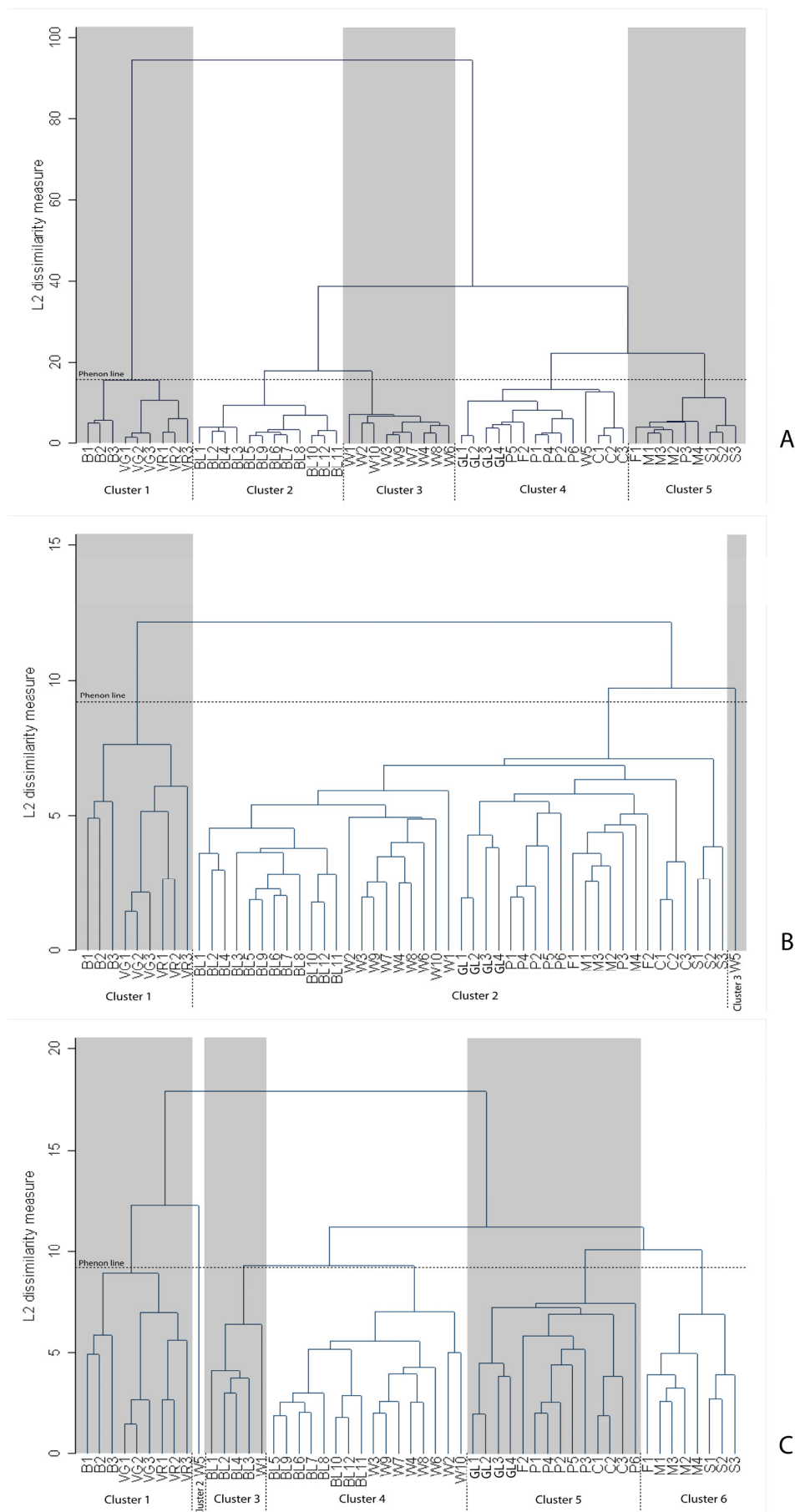


Figure 8.7: HCA dendrogram: variables.

HCA of the samples, using the Ward method, and with the phenon line drawn at a linkage distance of *ca.* 15, produces 5 clusters (Fig. 8.8a). Cluster 1, which comprises Bell Bridge sediments and the varved sediments from Blackhall Wood, is connected to the other clusters at an elevated Euclidean distance, thus indicating a geochemically distinctive suite of elements. Cluster 2 is made up of diamictons from Blackhall Wood, with individual stratigraphic units not distinguishable based on the position of the phenon line. The same is true for Willowford, with all but diamicton sample W5 assigned to cluster 3. Similarities between the geochemistry of samples at Willowford and Blackhall Wood are assumed because the clusters (2 and 3) are linked at a low Euclidean distance. Cluster 4 contains samples from Glasson, Plumpe Farm (except P3) and Carleton, plus samples W5 and F2. Cluster 5, which is linked to cluster 4 at a low Euclidean distance, consists of samples from

the Swarthy Hill and Maryport sites, plus samples P3 and F1. The low Euclidean distance indicates geochemical similarities between clusters 4 and 5.



< Figure 8.8: HCA dendograms with groupings (phenon line): (a) ward linkage; (b) average linkage; (c) complete linkage.

If the phenon line is moved to a linkage distance of  $\sim 9$ , individual stratigraphic units are revealed. This includes the summer and winter laminae of the varves at Blackhall Wood as well as all three diamicton units at Blackhall Wood (Fig. 8.8a) (Livingstone *et al.*, in press a). The average linkage method, with the phenon line at  $\sim 9$ , produces 3 clusters (Fig. 8.8b). Cluster 1 comprises the Bell Bridge and the varved sediments, whilst cluster 3 is diamicton sample W5 from Willowford. The rest of the samples are all assigned to cluster 2 (Fig. 8.8b). In contrast, the complete linkage method (with the phenon line at  $\sim 8$ ) produces 6 clusters (Fig. 8.8c). Cluster 1 is identical to both the Ward and average linkage methods, whilst cluster 2 contains the anomalous W5 sample (Fig. 8.8c). Cluster 3 contains the lower 4 diamicton samples from Blackhall Wood plus W1. Cluster 4, which is connected at a low linkage distance with cluster 3, comprises the rest of the Blackhall Wood and Willowford samples (Fig. 8.8c). Clusters 5 and 6 are almost identical to the respective clusters (4 and 5) identified by the Ward method (Fig. 8.8a, c).

#### 8.4.2 Particle size analysis

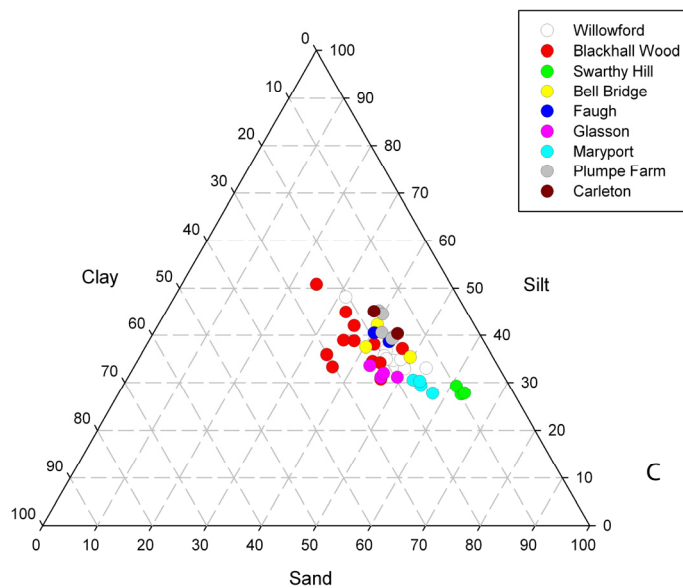
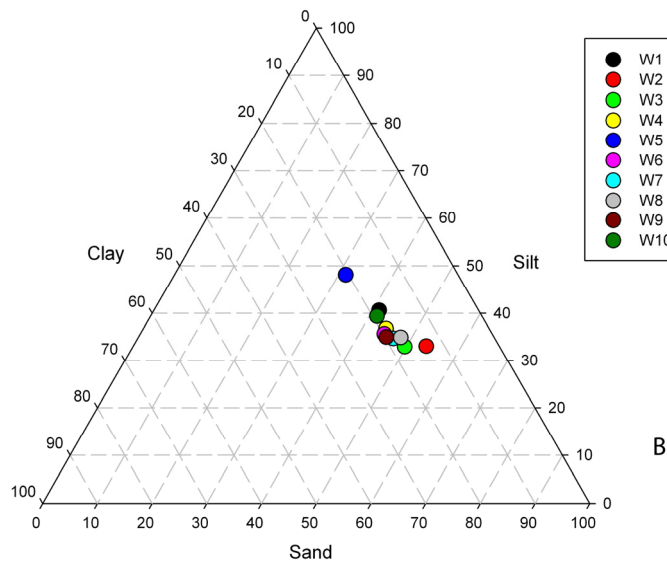
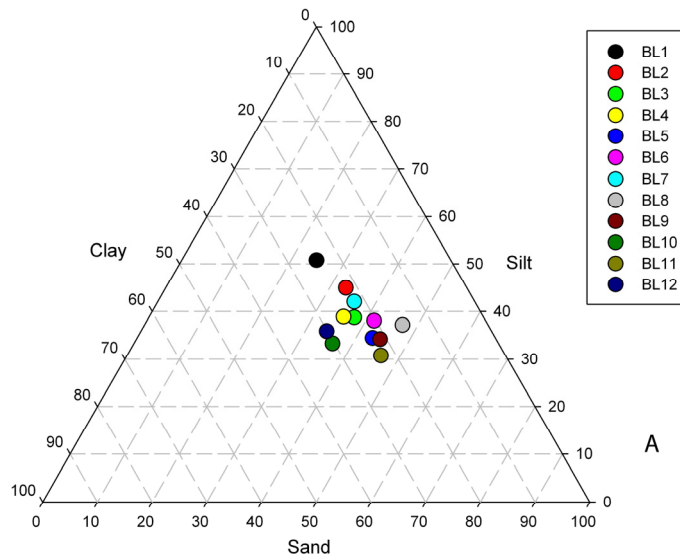
Particle size analysis on the diamictons at Blackhall Wood reveals a distribution consisting of 25 - 47% sand, 31 - 51% silt and 16 - 30% clay (Table 8.3). Although they are reasonably tightly grouped (Fig. 8.9a) a number of patterns are revealed on closer inspection. The upper diamicton unit, which overlies the varved sediments (Fig. 8.4), has in general a higher proportion of clay (Fig. 8.9a). This is in contrast to the lower diamicton unit, identified from the complete linkage and Ward approaches to HCA (Fig. 8.8a, c) which has elevated proportions of silt (Fig. 8.9a). The middle diamicton and debris-flow unit have similar particle size distributions characterised by high proportions of sand (Fig. 8.9a).

Field site	Sand (%)	Silt (%)	Clay (%)
<i>Willowford</i>	31 – 54	33 – 48	13 – 21
<i>Blackhall Wood</i>	25 – 47	31 – 51	16 – 30
<i>Swarthy Hill</i>	61 – 63	28 – 29	9 – 10
<i>Bell Bridge</i>	40 – 49	35 – 42	15 – 22
<i>Faugh</i>	40 – 44	39 – 40	17 – 19
<i>Glasson</i>	43 – 49	31 – 34	20 – 23
<i>Maryport</i>	52 – 57	28 – 31	15 – 17
<i>Plumpe Farm</i>	39 – 44	39 – 45	16 – 18
<i>Carleton</i>	38 – 45	40 – 45	15 – 17

Table 8.3: Particle size analysis: percentage range of the sand, silt and clay fractions for each field site.

Diamicton within the section at Willowford contains 31 - 54% sand, 33 - 48% silt and 13 - 21% clay (Table 8.3). The particle size analysis does not discriminate between the reddish and brownish-black diamictons (Fig. 8.4, 8.9b), although, like the HCA, W5 is easily distinguishable

(Fig. 8.9b). It is characterised by a lower proportion of sand and elevated levels of silt, compared to the rest of the samples (Fig. 8.9b).



< Figure 8.9: Ternary plots: particle size analysis: (a) Blackhall Wood; (b) Willowford; and (c) all sites.

A comparison of the particle size distributions of the diamicton samples at each location reveals a number of interesting groupings and patterns (Table 8.3, Fig. 8.9c). It can be observed that, apart from Blackhall Wood (BL) and Willowford (W), distinctive stratigraphic units at individual sites are not identifiable (Fig. 8.9c). Diamicton at Swarthy Hill, and to a lesser extent Maryport, are characterised by high proportions of sand compared to other sites (Table 8.3, Fig. 8.9c). The Faugh, Bell Bridge, Plumpe Farm and Carleton diamictons are characterised by similar particle size distributions, whilst those at Glasson have a relatively high percentage of the clay-sized grain fraction (Fig. 8.9c).

## 8.5 Discussion

Geochemical analysis has been used in collaboration with flow-phase and stratigraphic evidence to correlate stratigraphic relationships between disparate field sites and to delineate ice-flow histories for the region. An Event Stratigraphy has been compiled (Table 8.4) to summarise and explain the groupings identified within this, and previous Chapters.

Event	Flow-phases (Chapter 2)	Geochemical samples	Provenance	Basis for assigning samples to regional events and source provenances
Scottish Re-advance	SF1; LT6	PL1-6; C1-3	PL & C: Scottish.	PL: Stratigraphy and erratic content. C1-3: Geochemical similarity to PL and GL, and stratigraphic position.
Deglaciation				
Blackhall Wood Re-advance	LT5	BL10-12; GL1-4; S1-3; M1-4	BL: Lake District-Vale of Eden. GL: Scottish. M & S: Combined Scottish & Lake District-Vale of Eden.	Geomorphic association with flow-phase LT5.
Deglaciation		BL7-8; VR1-3; VG1-3	BL: Lake District-Vale of Eden. VR/VG: Local bedrock (Solway Lowlands-Vale of Eden).	Stratigraphic relationship, bracketed between tills from the Blackhall Wood Re-advance and Main Glaciation
Main Glaciation	LT1-3; ST1-3	W1-10; BL1-6, 9; B1-3	W: Combined Tyne Gap & Lake District-Vale of Eden. BL: Lake District-Vale of Eden. B: Local bedrock (Vale of Eden)	W: Geomorphology and stratigraphy/sedimentology. BL: Geochemical similarity to BL10-12, and stratigraphic position. B: Stratigraphy and sedimentology
Middle Sands				
Early Scottish Advance				



< Table 8.4: Updated Event Stratigraphy which illustrates the stratigraphic framework, provenance, related flow-phases and the evidence for assigning geochemical samples to particular flow-phases and source regions.

### 8.5.1 *Vertical stratigraphies*

Two of the nine fieldsites, Willowford and Blackhall Wood (Fig. 8.1) contain multiple units of diamicton, as defined by stratigraphic logging techniques (Fig. 8.4). The Blackhall Wood stratigraphy contains three diamictons, with the uppermost diamicton underlain by the varved sediments analysed in this study (Fig. 8.4) (cf. Livingstone *et al.* (in press a) for a thorough description of the glacial sediments and see Chapter 6). In general there is little difference between BL1-12, with samples occupying similar bivariate spaces in PCA, and both the Ward and average linkage methods of the HCA failing to distinguish the 3 diamicton units at this site. This geochemical similarity in the deposits indicates comparable ice-flow histories sourced from the same region (Table 8.4), in agreement with the clast-lithological analysis, macrofabric data and stratigraphic logging carried out by Livingstone *et al.* (in press a). Thus the lower diamicton units, which are not associated with a geomorphic signature, can be assigned to northwards flow out of the Vale of Eden (Table 8.4), in-line with flow-vector evidence (e.g. drumlin orientation) from the geochemically similar upper diamicton.

Despite the geochemical homogeneity demonstrated for the Blackhall Wood samples, the complete linkage method (HCA) successfully separates out the lower 4 diamicton samples (cluster 3), and all 3 diamicton units at this site are identified when the phenon line is lowered to *ca.* 9 (Ward method). Particle size analysis also reveals different proportions of sand, silt and clay within each unit (Fig. 8.9a). The ability of both HCA and particle size analysis to identify stratigraphic units, provides additional verification for the division of glacial sediments outlined in Fig. 8.4 and determined previously in Livingstone *et al.* (in press a) The high clay content of the ‘upper’ diamicton probably resulted from cannibalisation and incorporation of the underlying fine-grained varves (e.g. Hicock & Dreimanis, 1989; Hooyer & Iverson, 2000; Evans, *et al.*, 2006). The minor geochemical dissimilarities of the stratigraphic units could be explained by the positive relationship with component 1 and 2 of the PCA (Fig. 8.6a). This seems to indicate a gradual change up sequence, related to reductions in Li, V, Co, Ga, Sn, U, Fe (component 1), Ca, Ce, Mg, Sr and Y (component 2). Such a pattern could potentially result from the declining influence of local bedrock, partial cannibalisation of each unit, weathering or minor shifts in ice-flow.

The exposure at Willowford is made up of two interfingering diamictons, one brownish-black and one reddish (Fig. 8.4). These are described in detail by Livingstone *et al.* (in press b) and also in Chapter 4. PCA, HCA and particle size analysis all agree upon two groups: (1) diamicton sample W5; and (2) the rest of the samples (Figs. 8.6, 8.8 & 8.9b). Hence geochemical and particle size

analysis were both unable to discriminate between the reddish and brownish-black diamicton lithofacies, even in response to a lowered phenon line. This would suggest that despite the differences in colouration, both diamictons have similar provenances. As the reddish colouration is likely to be derived from the Permo-Triassic sandstones situated in the Solway Lowlands and Vale of Eden to the west of the Tyne Gap (Fig. 8.1), and the brownish-black colouration from Carboniferous rocks in the Tyne Gap itself (Trotter, 1929), it is logical to assume that both diamictons contain a mixed signature derived from both far-travelled and local lithologies. This theory is given further credence by the geochemical signature and particle size distribution of the diamicton sample W5. The partially brecciated rock, derived from the local Carboniferous bedrock (and slowly crushed and incorporated as diamicton (cf. Evans, *et al.*, 2006)), contains a different signature to the similarly coloured brownish-black diamicton, thereby demonstrating the difference between a locally-sourced diamicton (W5) and a cannibalised diamicton characterised by both a local and regional provenance signal. This interpretation demonstrates that the brownish-black diamicton was initially derived from local sources before being slowly incorporated into the matrix of the far-travelled diamicton.

Apart from Faugh, where the HCA predicts two groups, all other field sites are characterised by geochemically homogenous diamictons, which is as expected given the simple stratigraphy and geometry of these units (Fig. 8.4). This homogeneity has special resonance at Glasson, a 19 m high diamicton-cored drumlin, exposed along its stoss end (Figs. 8.1 & 8.4). Geochemical and particle size analysis, coupled with a bimodal macrofabric data mirroring the long profile of the drumlin (Fig. 8.10) and a macroscopically massive diamicton structure, suggests that the sediment formed synchronously during a relatively rapid period of deposition. The plausibility of such a rapid phase of drumlin construction in thick subglacial till has recently been demonstrated by observations on a modern ice-sheet bed by Smith *et al.* (2007) and King *et al.* (2009).

#### 8.5.2 *Stratigraphic framework for the Solway Lowlands-Tyne Gap*

The results presented in this study have enabled stratigraphic correlations to be made between diamicton units throughout the Solway Lowlands-Tyne Gap. These are summarised in Fig. 8.11 and Table 8.4, and discussed in detail below.

The Bell Bridge deposits and the varved sediments of Blackhall Wood comprise the most distinctive assemblage (1), with the various geochemical analyses displaying total compatibility (Fig. 8.11). Given that Bell Bridge lies only 8.5 km south of Blackhall Wood (Fig. 8.1) and within the same field of S-N orientated lineations (cf. Livingstone *et al.*, 2008) it is significant that the diamictons of Blackhall Wood display unique provenance signals. This suggests that the diamictons are associated with different ice-flow events, an entirely plausible explanation given that the diamicton at Bell Bridge is from an exposure in the bottom 4 m of a 40 m high lineation.

Alternatively, because the diamicton directly overlies a glacioteconite derived from the underlying Westphalian and Stephanian mudstone (Figs. 8.4 & 8.12), the geochemical signal could simply be a function of the local bedrock (cf. Boulton 1996a, b; Evans *et al.*, 2006).

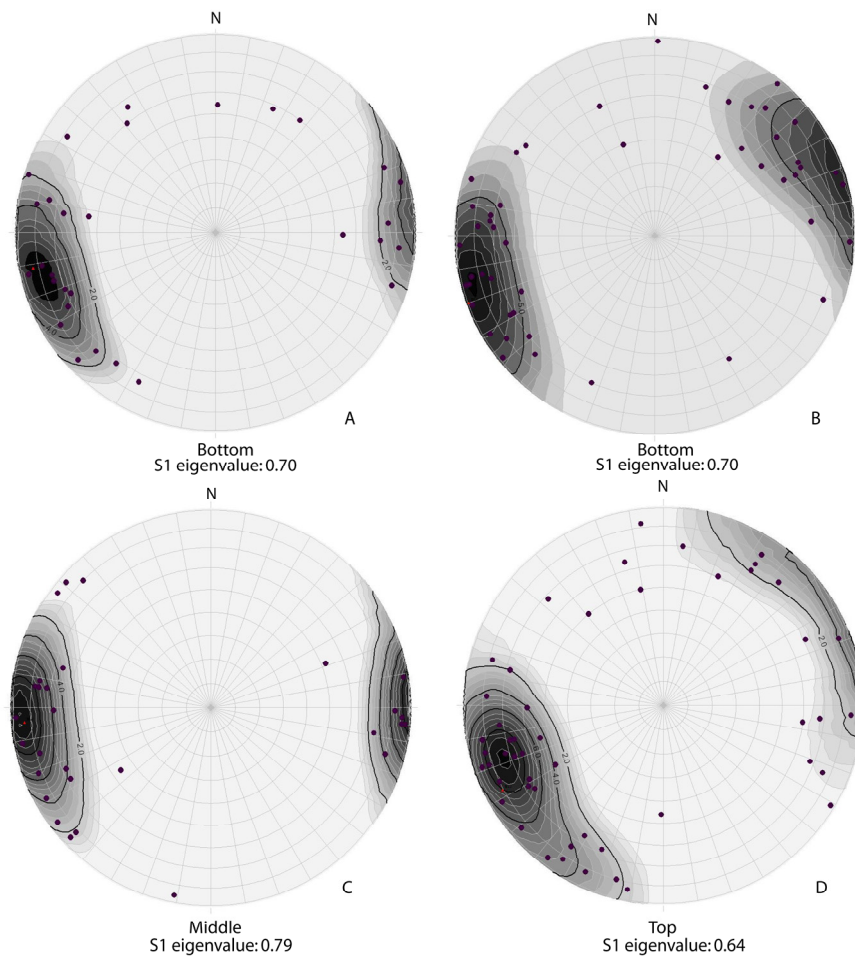


Figure 8.10: Macrofabric data for Glasston field site.

Assemblage 2 comprises the Maryport and Swarthy Hill deposits (Fig. 8.11), based on the HCA and particle-size analysis. This assemblage is to be expected given the geographical proximity of the two sites (Fig. 8.1) and their association with the same flow-set (LT5: cf. Livingstone *et al.*, 2008, in press a). The Plumpe Farm, Glasston and Carleton deposits make up the third assemblage (Fig. 8.11), based on the results of the PCA and HCA. The Plumpe Farm and Glasston deposits are associated with ice from Scotland (based on stratigraphic and geomorphic inferences: cf. Phillips *et al.*, 2007; Livingstone *et al.*, 2008), which implies that the diamicton at Carleton is also of Scottish origin (Table 8.4). This inference is merited because the diamicton displays many of the properties associated with the so called ‘Scottish re-advance till’ (cf. Trotter 1922, 1923, 1929), namely: (a) its geographic position is just to the SE of Carlisle; (b) it caps a series of deltaic and glaciofluvial deposits; and (c) it is only 0.4 m thick (Fig. 8.4). The clustering of both Glasston and Plumpe Farm diamictons demonstrates that although the geochemical analysis can pick out the Scottish provenance, it is unable to distinguish between the NE-SW (Glasston) and W-E (Plumpe Farm) moving ice-flow events which are defined by discrete ‘flow-sets’ associated with flow-phases LT5

and SF1 respectively (Table 8.4) (cf. Livingstone *et al.*, 2008). It is also significant that these ‘Scottish’ diamictos do not display the noticeably finer textures described by Huddart (1970). This misfit between two datasets probably stems from the variable influence of locally complex stratigraphies and geologically diverse bedrock which precludes regional generalisations about till provenance based on particle size distributions alone.

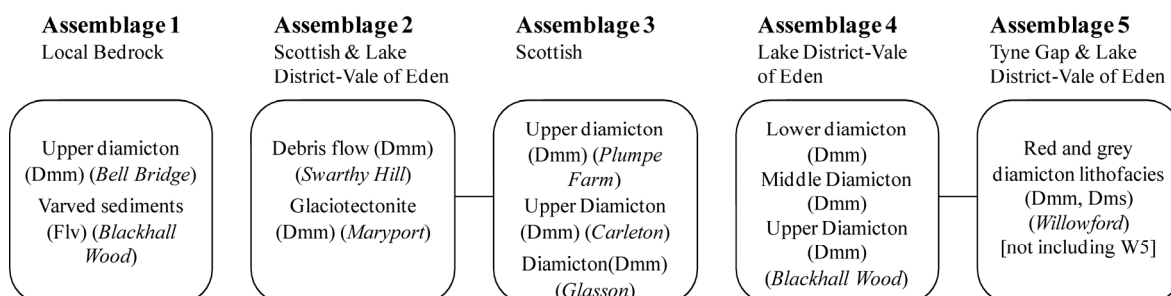


Figure 8.11: Proposed stratigraphic framework for the diamictos of NW England.

Assemblages 2 and 3 are linked at a low Euclidean distance (HCA, Fig. 8.8), implying some form of geochemical relationship. Because the diamictos of both Plumpe Farm and Glasson are sourced from Scotland (Livingstone *et al.*, 2008) and Lake District-Vale of Eden ice did not reach as far north as Plumpe Farm (cf. Evans, *et al.*, 2009), assemblage 2 must be partly associated with assemblage 3 (and not vice versa), suggesting that the diamictos of Swarthy Hill and Maryport have partially incorporated Scottish sediments (Fig. 8.11; Table 8.4). This could either relate to a convergence of Scottish and Lake District ice during draw-down into the Irish Sea Basin at a late phase of glaciation (Smith, 2002; Livingstone *et al.* 2008), or the cannibalisation and incorporation of an earlier Scottish advance, possibly associated with the retreat of ice prior to the Blackhall Wood re-advance (Livingstone *et al.*, in press a). It is difficult to comment on the diamicton at Faugh sand pit because the results of the statistical analysis are inconclusive. It could contain material either of a Scottish provenance or of a mixed Scottish and Lake District-Vale of Eden provenance.

Assemblage 4 consists of the Blackhall Wood diamictos (Fig. 8.11), based on the PCA and the HCA (Figs. 8.6 & 8.8). Despite belonging to the same flow-set (Livingstone *et al.*, 2008), there is poor correspondence between the diamictos of Swarthy Hill/Maryport and Blackhall Wood. This difference is attributed to the mixed provenance of assemblage 2, the complex bedrock geology of the region and the cannibalisation and incorporation of underlying sediments into the capping diamictos (Livingstone *et al.*, in press a). The diamictos of Blackhall Wood are linked at a low Euclidean distance with the diamictos of Willowford (assemblage 5) (Fig. 8.11). PCA, which also exhibits a close association between the two assemblages (Fig. 8.6), demonstrates high loading of component 2. One of the main elements associated with component 2 is Ca, which can be linked with confidence to the Carboniferous limestone (Fig. 8.2). This is not unexpected because Willowford is situated within the Carboniferous limestone of the Tyne Gap, whilst Blackhall Wood

must have been sourced from the Vale of Eden, which is fringed by Carboniferous limestone. Indeed, the lowermost diamicton of the Blackhall Wood sequence is interpreted by Livingstone *et al.* (in press a) to have converged on the Tyne Gap (flow-phases LT1-3: Chapter 2), thus indicating a common flow-direction during an early phase of ice movement (Table 8.4).



Figure 8.12: Photograph: mosaic of the glaciotectionite and upper diamicton at Bell Bridge showing the influence of local bedrock on till formation.

## 8.6 Conclusions:

This study has highlighted the potential of geochemical analysis in constraining ice-flow histories (Table 8.4) within a geologically and glacially complex region of the BIIS. A number of important conclusions have arisen from this quantitative approach to glacial stratigraphy:

### 8.6.1 Statistical approach to stratigraphic correlations:

- i. HCA and PCA techniques were successfully employed to distinguish and therefore verify stratigraphic units previously identified independently using geomorphological mapping, stratigraphic logging, clast lithological analysis, macrofabric analysis and micromorphology.
- ii. The statistical approach to geochemical investigation is an exploratory one in the UK setting and therefore should be used either to formulate hypotheses or as part of a suite of techniques until a denser network of samples is available for the region and clear bedrock sources can be assigned to particular geochemical signatures. These developments will enable the reconstruction of glacial flow patterns to a level compatible with that achieved in North America and Scandinavia.

### 8.6.2 Ice-flow history of the Solway Lowlands-Tyne Gap:

- i. Five major geochemical assemblages are identified within the central sector of the BHS (Fig. 8.11; Table 8.4), assigned to Scottish, Scottish and Lake District-Vale of Eden, Tyne Gap and Lake District-Vale of Eden, Lake District-Vale of Eden, and local bedrock provenances. In particular this approach has: (a) revealed the Carleton diamicton to be a Scottish re-advance till; (b) identified the existence of Scottish ice along the west coast of Cumbria either before or during the Blackhall Wood re-advance; (c) revoked the existence of a Scottish diamicton pertaining to 'Early Glaciation' at Blackhall Wood; (d) identified three distinctive diamicton units at Blackhall Wood; (e) recognized the similar provenances of both the brownish-black and reddish diamictons at Willowford, thus indicating a mix of local and regional lithologies; and (f) demonstrated that the varves at Blackhall Wood associated with 'Blackhall Wood' glacial Lake were sourced from the same lithology as Bell Bridge, thus indicating deposition from the south.
- ii. The local bedrock exerts a strong influence at some sites (Bell Bridge), particularly where there is evidence of bedrock being rafted up into the local diamictons (Bell Bridge and Willowford). Indeed it is essential in geochemical investigations to separate out the local and regional signal. This study was restricted in this regard because it adopted a relative approach, whereby clusters of elements were not assigned to individual geologies owing to the lack of data available on the geochemistry of non-glacial regolith.
- iii. Particle size analysis does not identify regional trends but rather localised signatures, which are a function of the underlying sedimentology and geology (underscoring the dangers of regional till correlation on the basis of PSA alone). This is demonstrated by the clayey upper diamicton at Blackhall Wood capping the varved sediments, and the sandy diamictons at Swarthy Hill and Maryport which have been associated with the glaciotectionisation of a series of glaciofluvial and glaciolacustrine sediments.

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## Chapter 9

# Glacial history of the central sector of the last British-Irish Ice Sheet

## 9.1 Introduction

This chapter presents an overview of the glacial history of the central sector of the last British-Irish Ice Sheet (BIIS) (summarised in Table 9.1). This reconstruction is based on the preceding eight chapters and involves data collected using a range of geomorphological, sedimentological and geochemical techniques. Six main stages have been identified, of which stage I is interpreted to encompass the period of maximum ice expansion, while stages II-VI relate to the deglacial history of the region. Also discussed is a pre-stage I event associated with the expansion of ice out of upland regions (Boulton & Hagdorn, 2006; Hubbard *et al.*, 2009), and the evidence for an 'Early Advance' of Scottish ice up the Vale of Eden (Trotter, 1929). The central sector of the last BIIS is shown to exhibit dynamic ice-flow phasing characterised by flow switches, initiation (and termination) of ice streams, draw-down of ice into ice streams, repeated ice-marginal fluctuations and the production of large volumes of meltwater, often impounded to form ice-dammed glacial lakes. A major ice-stream has been recognised emanating eastwards through the Tyne Gap, a topographic low situated between upland massifs. However, the Irish Sea Ice Stream acted independently of the central sector of the BIIS during stage I probably due to the emplacement of an ice-divide in the northern sector of the Irish Sea Basin. Two major oscillations have been identified within the central sector of the BIIS; the Blackhall Wood Re-advance and Scottish Re-advance. During the Blackhall Wood Re-advance ice streamed SW into the Irish Sea Basin and at this stage must have acted as a tributary to the Irish Sea Ice Stream. The relevance of these ice-flow stages to the general glacial history of the BIIS is discussed and, where possible, the stages are correlated with known glacial events (see Table 9.1).

Event	Stage	Key features	Diamicton samples	Flow-phases	Regional Correlation	Date	Possible triggers
Scottish Re-advance	VI	Holme St. Cuthbert delta	PL1-6; C1-3	SF1; LT6; (LT7)	Killard Point Stadial	~16.8	Heinrich Event 1
Deglaciation	V	Glacial-Lake Carlisle; Wiza Beck mwc; Brampton kame belt; Pennine escarpment mwc					
Blackhall Wood Re-advance	IV	Ice-flow tributary of the Irish Sea Ice Stream; Brampton kame belt	BL10-12; GL1-4; S1-3; M1-4	LT5	Gosforth Oscillation; Clogher Head Stadial (?); SW ice-flow over the Isle of Man	~19.5	19 ka BP meltwater pulse
Deglaciation	III	Glacial-Lake Blackhall Wood; Brampton kame belt	BL7-8; clastic varves	(EC1; LT4)	Glacial-Lake Wear		
Main Glaciation	II	SE flow down the N. Tyne Valley	W1-10; BL1-6, 9; B1-3	EC1; LT4; ST3-4; ice flow down the Vale of Eden	Rapid retreat of Irish Sea Ice Stream from Celtic Basin		
	I	Eastward ice-flow through Tyne Gap Ice Stream and Stainmore Gap		LT1-3; ST1-2; ES1	Maximum extent of Irish Sea Ice Stream and Irish Ice Sheet	~23-24	
Middle Sands	Pre-stage I						
Early Scottish Advance	(?)						



< Table 9.1: Event Stratigraphy for the central sector of the last British-Irish Ice Sheet. Key features correlate sediment-landform assemblages to discrete stages of the Event Stratigraphy (mwc is an abbreviation of meltwater channels). Diamicton samples refer to the Event Stratigraphy compiled in Chapter 8 and indicate where the stratigraphy fits into the overall flow-phase history. Flow-phases refer to geomorphological mapping summarised in Chapter 2 and utilised in chapters 3-8. All dates are in calibrated years and question marks indicate some degree of uncertainty.

## 9.2 Overview of the glacial history

### 9.2.1 *Pre-Stage I*

#### Ice expansion out of upland areas:

A ubiquitous mode of ice flow implicit within previous glacial reconstructions has been the infringement of Scottish ice up the Vale of Eden and into the Stainmore Gap (533 m) (Trotter, 1929; Hollingworth, 1931; Trotter & Hollingworth, 1932; Huddart, 1970; Catt, 1991). Evidence for this flow phase is limited to an extensive suite of Scottish erratics and several exposures of lower red-brown till (Gillcambon Till Formation) interpreted to have been deposited during this advance (Trotter, 1929; Hollingworth, 1931; Trotter & Hollingworth, 1932; Huddart, 1970). However, stratigraphic and geochemical investigations in the northern sector of the Vale of Eden fail to discriminate between lower and upper diamicton lithofacies (Livingstone *et al.*, in press a; Chapter 8), while numerical ice-sheet modelling results (Evans *et al.*, 2009) fail to reconcile such a southerly advance of Scottish ice. A glaciologically more plausible explanation is that the encroachment of Scottish ice into NW England during a previous glaciation, and/or an early stage of the Late Devensian glaciation provided a ready supply of Southern Upland erratics which were subsequently dispersed by localised ice flows (cf. Evans *et al.*, 2009).

If this strong flow of Scottish ice into NW England occurred during the last glaciation then the prevalence of Scottish erratics throughout the stratigraphic sequence demand that it be associated with an early ice movement (pre-stage I). Such a movement has been identified in the NE Irish Sea Basin, with Salt and Evans (2004) and Roberts *et al.* (2007) both recognising that initial ice flow into the region was driven by Highland ice sourced in the Firth of Clyde (also see Boulton & Hagdorn, 2006). This ice flow resulted in the build up of ice in the northern sector of the Irish Sea Basin eventually leading to the southward migration of the ice divide (Boulton & Hagdorn, 2006). Within the central sector of the BIIS this ice-flow stage was likely characterised by congestion and thickening of ice throughout the Vale of Eden and Solway Lowlands reinforced by the expansion of ice out of the Lake District, Howgill Fells and northern Pennines. This interpretation is based upon *a priori* knowledge governing the general growth and expansion of ice sheets (instantaneous glacierization: Ives *et al.*, 1975).

### 9.2.2 *Stage I*

#### Eastwards ice flow through prominent topographic corridors of the north Pennines:

During stage I the ice divide straddling the northern sector of the Irish Sea Basin reached its southern-most point, pinning the Southern Upland and Lake District ice masses against the Cumbrian coast and forcing flow eastwards (cf. Evans *et al.*, 2009; Fig. 9.1). Ice in the central sector of the BIIS had also reached its maximum observed thickness overwhelming much of the region and moving independently of topography over cols in the northern Pennines (see Fig. 9.1). The highest mountains in the north Pennines, Cold Fell and Cross Fell, maintained local ice caps which fed into the main body of ice (Trotter, 1929; Vincent, 1969; Mitchell, 2007). These local ice-caps on the highest massifs are interpreted as cold-bedded ice-plateaus, as inferred by their interfluvial geomorphology which has survived glaciation (Kleman & Glasser, 2007), and the paucity of subglacial bedforms indicative of temperate ice flow (Mitchell, 2007). Stage I therefore comprises the greatest recorded mass of ice in the central sector of the BIIS and is thus correlated with maximum ice sheet expansion in the Irish Sea Basin at ~23-24 ka BP (Table 9.1). It is at this time that the grounded Irish Sea Ice Stream reached its maximum extent, advancing into the Celtic Basin and reaching as far south as the Isles of Scilly (Scourse *et al.*, 1990; Scourse, 1991; Hiemstra *et al.*, 2005; Ó Cofaigh & Evans, 2007). This is supported by 26 reworked shells from Irish Sea till from the southern coast of Ireland which have AMS  $^{14}\text{C}$  dates between 23.9 and 24.6 cal. ka BP (Ó Cofaigh & Evans, 2007). In addition, the off-shore record indicates a peak in ice rafted debris from Goban Spur, off the Celtic Basin, at ~24-25 ka BP (Scourse *et al.*, 2009), whilst Greenwood & Clark (2009) constrain the maximum limits of the Irish Ice Sheet to between ~28-24 ka BP (stage III-IV of Greenwood & Clark, 2009). Therefore, the central sector of the BIIS did not function as a tributary of the Irish Sea Ice Stream during its maximum expansion to the Scilly Isles. The glacial model for the Irish Ice Sheet proposed by Greenwood and Clark (2009) envisages a broad ice-divide centred over the northern Irish Sea Basin during maximum ice extent, thus corresponding to, and constraining the geomorphological evidence observed in this thesis.

The build up of ice in the central sector of the BIIS coincided with northwards migration of the Vale of Eden ice divide (Letzer, 1978). This is reconciled with geomorphological evidence indicating convergent flow into and across Stainmore Gap (533 m) from the NW, W and SW (ice-flow phases ES1, ST1: Chapter 2) (Hollingworth, 1931; Letzer, 1978; Livingstone *et al.*, 2008; Fig. 9.1). Erratic trains of the distinctive Shap granite and Permian Brockram (Trotter, 1929; Hollingworth, 1931) constrain the easterly movement of Lake District ice. Flow from the Howgill Fells and the eastern sector of the Lake District requires the emplacement of an ice divide straddling the two upland regions (Fig. 9.1). The convergence and flow of ice through the Stainmore Gap suggests that it acted as a major flow artery, transferring ice to the eastern side of

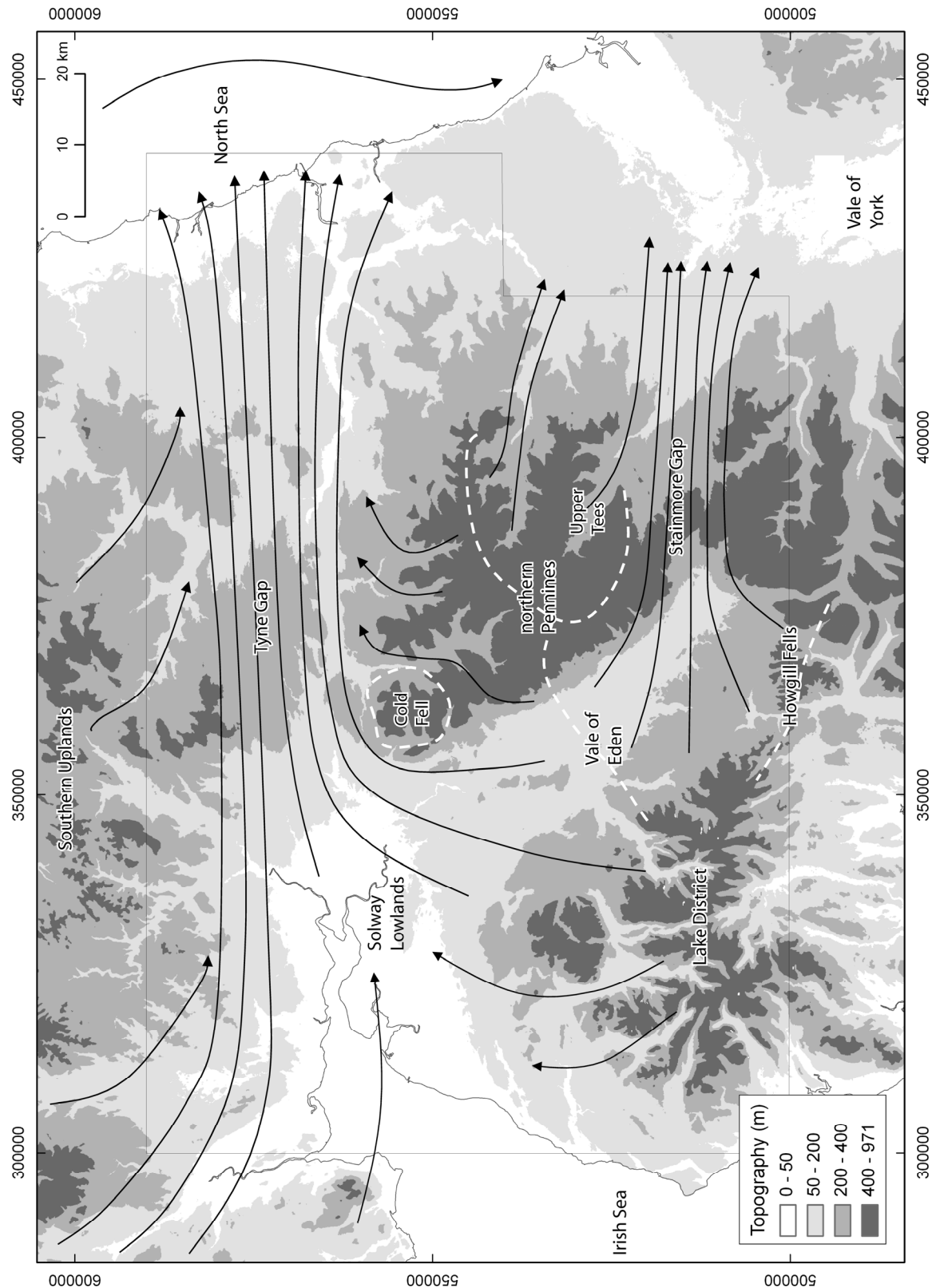
the country. Flow was supplemented by the Tees Glacier which provided a local outlet from the Cross Fell ice cap (cf. Mitchell, 2007; Fig. 9.1). The eastwards flow of ice through Stainmore Gap has been traced into the Vale of York where it formed a major trunk glacier which flowed southwards, depositing a reddish brown till and the Escrick and York morainic ridges (stage I and early in stage II; Table 9.1) (cf. Catt, 2007). The southerly limit of this trunk glacier is contentious, with Gaunt (1976, 1981) suggesting that it may have surged beyond the Escrick moraine into Lake Humber (which was impounded by the North Sea ice lobe) and to a temporary ice limit 50 km south of this moraine, marked by a discontinuous line of gravels at Wroot and Thorne (cf. Catt, 2007 for a review). Lake Humber is thought to have existed sometime between ~16.6 – 24 cal. ka BP (Gaunt, 1974; Bateman *et al.*, 2007).

Overprinted relationships of subglacial lineations in the Stainmore Gap (Livingstone *et al.*, 2008) reveal a shift in the influence of Ice Dispersal Centres (IDC) probably related to the southwards migration of the Vale of Eden ice divide (Letzer, 1978, 1987). As the influence of ice from the Vale of Eden waned, the Howgill Fells became the dominant IDC, resulting in NE ice flow over Stainmore Gap (ice-flow phase ST2: Chapter 2) (Livingstone *et al.*, 2008; Fig. 9.1). It can be speculated that southwards migration of the Vale of Eden ice divide towards the Howgill Fells and the subsequent expansion of ice from this region acted to deflect ice from the eastern sector of the Lake District (Shap region) northwards, nourishing the flow of ice across Cold Fell col and Tyne Gap. The shifts in ice flow trajectories driven by migrating ice divides and IDCs during this stage are envisaged to have played a key role in controlling the dynamics and influence of individual flow corridors across the Pennines.

During stage I, ice flow through the Tyne Gap (152 m) consisted of an eastwards moving topographic ice stream (Livingstone *et al.*, in press b; Fig. 9.1). It was characterised by convergent flow from the Scottish Southern Uplands and Lake District (see Chapter 8), with tributaries of locally sourced ice flowing into the main artery from the northern Pennines (Fig. 9.1). The sensitivity of the Tyne Gap ice stream to the migration of ice divides and IDCs is demonstrated by shifts in ice flow direction from NE to E (ice-flow phases LT1-3: Chapter 2) (Livingstone *et al.*, 2008, in press b; Fig. 9.1). These shifts reflect a change in dominance between the contribution of Lake District and Scottish Southern Upland ice to the regional flow through the Tyne Gap, and also a reversal in the migration of the Irish Sea Basin ice divide back towards Scotland (Livingstone *et al.*, in press b). The influence of the Tyne Gap ice stream dominated its immediate vicinity such that the Tweed-Cheviot ice stream was deflected further out into the North Sea relative to later flow stages (Raistrick, 1931; Livingstone *et al.*, in press b).

It can be speculated that the development of a fast flowing ice stream in the Tyne Gap acted to draw-down ice from surrounding feeder zones and tributaries (Price *et al.*, 2008). This provides a glaciological explanation for the migration of the Vale of Eden ice divide away from the zone of

convergence in the Tyne Gap, and the development and capture of flow across Cold Fell col and out of the Lake District (Fig. 9.1). Implicit within this reconstruction is that Stainmore Gap acted as a subsidiary ice flow artery and was therefore incapable of maintaining its northerly feeder zones under direct competition with the Tyne Gap ice stream.



< Figure 9.1: Stage I: Eastwards ice flow through prominent topographic corridors of the north Pennines. The grey box outlines the study area, while the white dotted lines refer to ice divides and the arrows indicate ice flow directions. The ice divide in the Vale of Eden is inferred from lineation orientations leading into the Stainmore Gap. Although this diagram depicts one moment in time, during stage I both the Stainmore Gap and Tyne Gap were associated with shifting ice flow directions in response to migrating ice divides and Ice Dispersal Centres. It must be noted that for all the Figures in this chapter, the absence of ice flow lines outside of the grey box does not implicate a lack of ice, rather a lack of knowledge regarding the direction and extent of ice flow during that particular stage.

### 9.2.3 *Stage II*

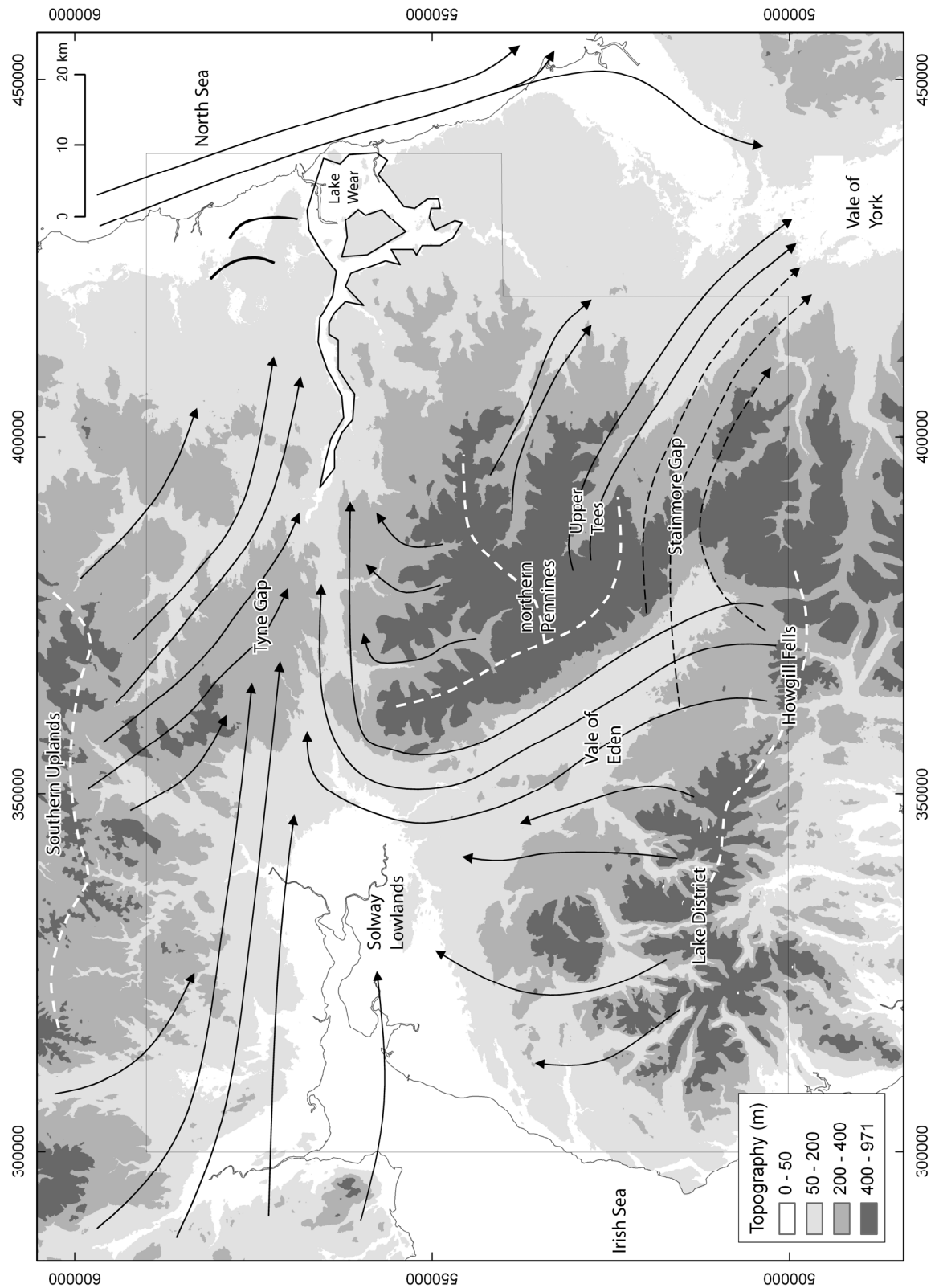
#### Cut-off of the Stainmore Gap as an ice through-flow and northwards migration of the Solway Firth ice divide:

By stage II, ice divides had started to migrate back into upland regions leading to significant changes in ice dynamics primarily related to the eastwards flow of ice through the Tyne and Stainmore Gaps. These two major ice-flow pathways are treated independently of each other during stage II as cross-cutting relationships fail to distinguish between the relative terminations of ice flow over each of these cols. However, given the height of the Stainmore Gap (533 m) compared to the Tyne Gap (152 m), and their relative locations compared to major IDCs and ice divides (Fig. 9.2), it is logically assumed that the Stainmore Gap would be more sensitive to changes in ice flow dynamics.

#### 9.2.3.1 *Stainmore Gap:*

As the ice divide originally situated in the Vale of Eden continued to migrate southwards into the Howgill Fells ice flow across Stainmore Gap from the western side of the Pennines began to weaken. This is demonstrated by the youngest set of subglacially-formed lineations of the mountain pass, which indicate SE flow of ice towards the Vale of York, dominated by ice sourced in the Upper Tees (ice-flow phases ST3-4: Chapter 2) (Mitchell, 2007; Livingstone *et al.*, 2008; Fig. 9.2; Table 9.1). Further retreat of the ice divide and surface lowering west of the Stainmore Gap eventually resulted in a switch in ice direction towards the north (Fig. 9.2), thereby severing the flow of ice over Stainmore col (Table 9.1) (Trotter, 1929; Letzer, 1987; Mitchell & Clark, 1994; Smith, 2002; Livingstone *et al.*, 2008; Evans *et al.*, 2009). This is clearly represented by cross-cutting patterns in the Vale of Eden with W-E orientated drumlins preserved in Stainmore becoming increasingly rare towards the west where the later northwards flow exerted a greater influence in re-moulding the landscape (cf. Evans *et al.*, 2009). Further changes in ice configuration are limited to the head of Tees Glacier, in a region over 600 m O.D., with ice flow shifting from a restricted N-

S (ice-flow phase ST3: Chapter 2) movement down the valley, to an easterly trajectory (ice-flow phase ST4: Chapter 2) as the ice centre expanded westwards (Fig. 9.2) (also see Mitchell, 2007).





< Figure 9.2: Stage II: Cut-off of the Stainmore Gap as an ice through-flow and northwards migration of the Solway Firth ice divide. The dashed arrows refer to an intermediate ice flow phase during the southerly migration of the Vale of Eden ice divide prior to the complete severance of the Stainmore Gap as an ice flow pathway. The thick black lines refer to still-stand positions of the ice front during westerly recession out of the Tyne Gap, which eventually resulted in the formation of Glacial Lake Wear (black polygon).

#### 9.2.3.2 *Tyne Gap:*

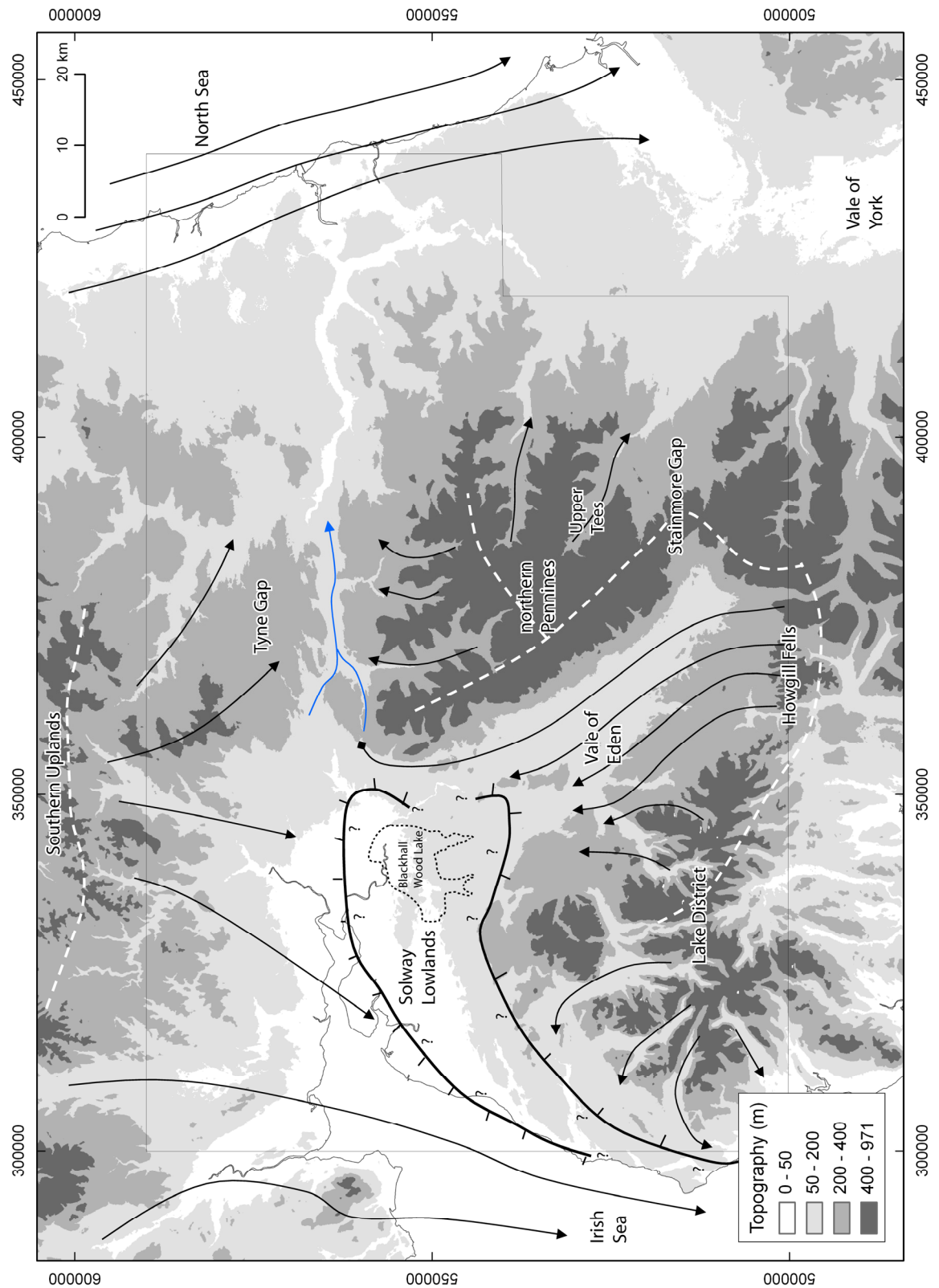
As the ice divide in the Irish Sea Basin continued to retract into the Southern Uplands, so the influence of the Tyne Gap as an easterly flowing ice stream weakened. Instead ice flow became dominated by Scottish ice moving SE down the North Tyne Valley and out of Bewcastle Fells, consequently (ice-flow phase LT4: Chapter 2; Table 9.1) deflecting the eastwards flowing ice artery (Livingstone *et al.*, in press b [stage II]; Fig. 9.2). Stage II is therefore associated with the eastwards migration of the Southern Uplands ice divide, first identified by Salt and Evans (2004) (stages D-F), causing the central Southern Uplands to become increasingly important as an IDC. This resulted in the gradual stagnation and incremental recession of the easterly margin, as delimited by a series of transverse moraines (Smythe, 1912), meltwater channels and kamiform deposits (Yorke *et al.*, 2007) at still-stand positions (Fig. 9.2). The gradual retreat of ice out of the eastern zone of the Tyne Gap resulted in the proto-formation of Glacial-Lake Wear dammed-up by the southerly flow of Tweed-Cheviot ice in the North Sea Basin (Teasdale & Hughes, 1999; Livingstone *et al.*, in press b).

#### 9.2.4 *Stage III*

Westerly retreat of ice from the Tyne Gap and the development of an ice free enclave within the Solway Lowlands:

Stage III was characterised by widespread deglaciation of the central sector of the BIIS (Table 9.1). The Tyne Gap ice stream continued to downwaste and retreat westwards, with meltwater routed into Glacial-Lake Wear via a major proglacial drainage network in the South Tyne Valley (Livingstone *et al.*, in press b; Fig. 9.3). Without the influence of the Tyne Gap ice stream, stages III-IV became associated with a more dominant Tweed-Cheviot ice stream, which was able to push inland from the North Sea (cf. Catt, 1991; Livingstone *et al.*, in press b). However, ice flow down the North Tyne Valley may have persisted during this (and the following) stage(s) (Table 9.1), while glacial flow from local Pennine IDCs may similarly have lingered (Fig. 9.3). As ice retreated across the Irthing-South Tyne watershed onto the reverse slope of the Tyne Gap it bifurcated into two distinct ice lobes, with Lake District ice receding SW into the Vale of Eden and Scottish ice

contracting west towards Dumfries-shire (Trotter, 1929; Fig. 9.3).



< Figure 9.3: Stage III: Westerly retreat of ice out of the Tyne Gap and the development of an ice-free enclave within the Solway Lowlands. The thick black lines refer to inferred ice fronts during the retreat of ice out of the Solway Lowlands, while the blue arrow depicts the proglacial drainage network that developed in the North Tyne Valley fed by stagnant ice in the Brampton region (black line with diamond head). The minimum extent of Blackhall Wood Lake is also tentatively illustrated, based on glaciolacustrine deposits in the region.

Further deglaciation as recorded by extensive spreads of glaciofluvial and glaciolacustrine sediments and debris-flow deposits mark the onset of an ice-free environment within the Solway Lowlands (Livingstone *et al.*, in press a; Fig. 9.3). Its western margin, off the Cumbrian coast, was delimited by the Irish Sea Ice Stream (Fig. 9.3), which, towards the end of stage III, acted to impede meltwater drainage, leading to the development of a proglacial lake in the Solway Lowlands (Livingstone *et al.*, in press a; Fig. 9.3). Clastic varves identified at Blackhall Wood indicate that the proglacial lake existed for at least 261 years, while stratigraphic correlation with other laminated sediments suggest that the lake occupied an area of at least 140 km<sup>2</sup> (Livingstone *et al.*, in press a; Fig. 9.3; Table 9.1). By stage III the Irish Sea Ice Stream had retreated out of the Celtic Basin in accord with the deglacial pattern exhibited by the central sector of the BIIS (Ballantyne, 2009). This is supported by dates which suggest that the ice stream rapidly retreated from its maximum limits (Table 9.1) to a position north of the Wicklow Mountains and western tip of Anglesey by ~21-19 cal. ka BP (cf. Ballantyne, 2009). However, it is still envisaged to have stretched down the west Cumbrian coast impounding lakes against higher ground in the Sellafield District, west Cumbria (Merritt & Auton, 2000) and reaching at least as far south as the Isle of Man (Thomas *et al.*, 2004; Roberts *et al.*, 2007). This is supported, in particular, by Roberts and co-authors (2007) who propose a two-phase model to explain the glacial geomorphic signature on the Isle of Man; an initial SE flow from the Southern Uplands (phase I) followed by SW flow out of the Solway Firth (phase II). Both phases are attributed to streaming associated with the Irish Sea Ice Stream and were therefore attributed to the LGM (Roberts *et al.*, 2007). Based on the Event Stratigraphy constructed for the central sector of the BIIS (Table 9.1), phase I of Roberts *et al.* (2007) can be confidently correlated with stage I and II in-line with evidence presented in this thesis, which suggests that the Irish Sea Ice Stream operated independently of the central sector of the BIIS during the LGM (Table 9.1). The only significant flow of ice into the Irish Sea Basin from the Solway Lowlands occurred during stage IV (ice-flow phase LT5: Chapter 2), during the Blackhall Wood Re-advance, and must therefore correlate with this event (see below). Therefore, ice is interpreted to have overran the Isle of Man throughout stages I to IV.

#### 9.2.5 Stage IV

##### Blackhall Wood Re-advance:

The overall pattern of retreat in the central sector of the BIIS was reversed during the Blackhall Wood Re-advance phase (stage IV). This re-advance was characterised by a switch in flow direction, with ice moving down the Vale of Eden before wrapping westwards around the northern edge of the Lake District and out into the Irish Sea Basin (ice-flow phase LT5: Chapter 2) (Livingstone *et al.*, 2008, in press a; Fig. 9.4). Subglacial lineations of Scottish provenance provide evidence for a coeval advance of Southern Uplands ice out of the Esk Valley, across the Solway Firth and down the Cumbrian coast (Livingstone *et al.*, 2008, in press a; Chapter 8; Fig. 9.4), with the SW flow direction indicative of a dominant ice divide in the central Southern Uplands, maintained from the previous stage. The arcuate 'swarm' of subglacial lineations (Fig. 9.4) display sedimentological and morphological characteristics typical of a fast-flow signature such as high elongation ratios (12:1) and evidence of pervasive glaciotectionic deformation (Stokes & Clark, 2001). It is envisaged that ice moving rapidly out into the Irish Sea Basin from the Solway Lowlands coalesced with, and functioned as a tributary of, the Irish Sea Ice Stream during this stage (Livingstone *et al.*, in press a). The fast flow signature of stage IV subglacial lineations in the Solway Lowlands suggests that ice was being vigorously drawn-down into the Irish Sea Ice Stream (Eyles & McCabe, 1989; Evans & Ó Cofaigh, 2003; Ó Cofaigh & Evans, 2007). A northwards transition into hummocky terrain and ribbed moraine delimits the lateral margin, where inferred ice velocities were significantly reduced (Livingstone *et al.*, 2008, in press a; Fig. 9.4). This transition could also have been at least partially influenced by topography, with the ribbed moraine separated from the highly elongate lineations by the Lias Plateau, which stands up to 50 m O.D. above the surrounding landscape.

The Blackhall Wood Re-advance (Livingstone *et al.*, in press a) is tentatively correlated with the Gosforth Oscillation in western Cumbria (~19.5 cal. ka BP) (Merritt & Auton, 2000). It was characterised by temporary retreat of Lake District valley glaciers and proglacial lake formation dammed against Irish Sea Ice (stage III). The subsequent Gosforth Re-advance (~19.5 cal. ka BP) (stage IV, Table 9.1) saw Lake District is re-coalesce with the Irish Sea Ice Stream leading to extensive drumlinisation of the region (Merritt & Auton, 2000). Masking of the arcuate lineation pattern (flow-phase LT5: Chapter 2) by the Holme St. Cuthbert deltaic sequence (Huddart, 1970; Chapter 7) verifies that this re-advance occurred prior to the Scottish Re-advance (Livingstone *et al.*, in press a; Chapter 7; Table 9.1). The influence of the central sector of the BIIS on the Irish Sea Ice Stream during this late stage (IV) of glaciation is replicated in the regional geological record with overprinted bedforms on the Isle of Man revealing a late SW phase of ice flow initiated in the Solway Lowlands (Roberts *et al.*, 2007: phase II). Cosmogenic dates from North Wales and eastern Ireland provide additional support for the occupation of the northern sector of Irish Sea Basin by the Irish Sea Ice Stream at ~19.5 cal. ka BP (Ballantyne, 2009).

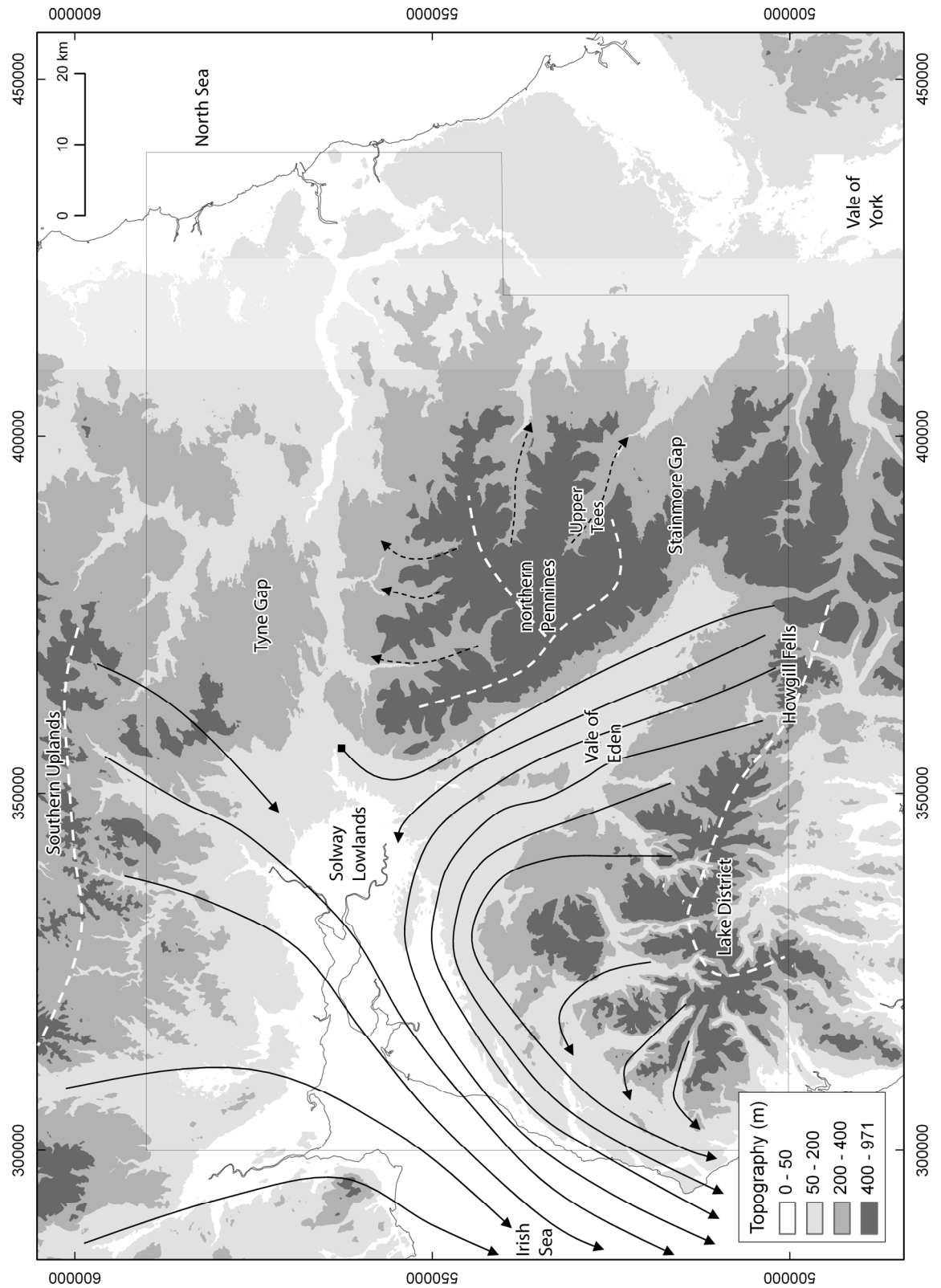


Figure 9.4: Stage IV: Blackhall Wood Re-advance. Dashed arrows in the northern Pennines illustrate the uncertainty associated with the continued existence of major outlet glaciers emanating from this ice cap. Note the continued existence of stagnant ice in the Brampton region.

The regional-scale of the Blackhall Wood-Gosforth Oscillation (~19.5 cal. ka BP) implies that it was triggered by an external forcing mechanism. This interpretation is given credence by its close chronological association with the 19 cal. ka BP meltwater pulse (Table 9.1), an event which caused an abrupt decrease in the Atlantic meridional overturning circulation leading to widespread cooling of the NE Atlantic (Yokoyama *et al.*, 2000; Clark, *et al.*, 2004; Hall *et al.*, 2006). This model is in general accord with recent findings in NE Ireland that elucidate two major re-advances during the deglaciation of the Late Devensian Irish Ice Sheet; the Clogher Head Stadial, between 18.3 – 17.0 cal. ka BP, and the Killard Point Stadial after 17.0 cal. ka BP (McCabe, *et al.*, 2005, 2007) (Table 9.1). These offer possible pan-Irish Sea correlations with the Blackhall Wood-Gosforth Oscillation (stage IV) and Scottish Re-advance (stage VI) respectively (Table 9.1).

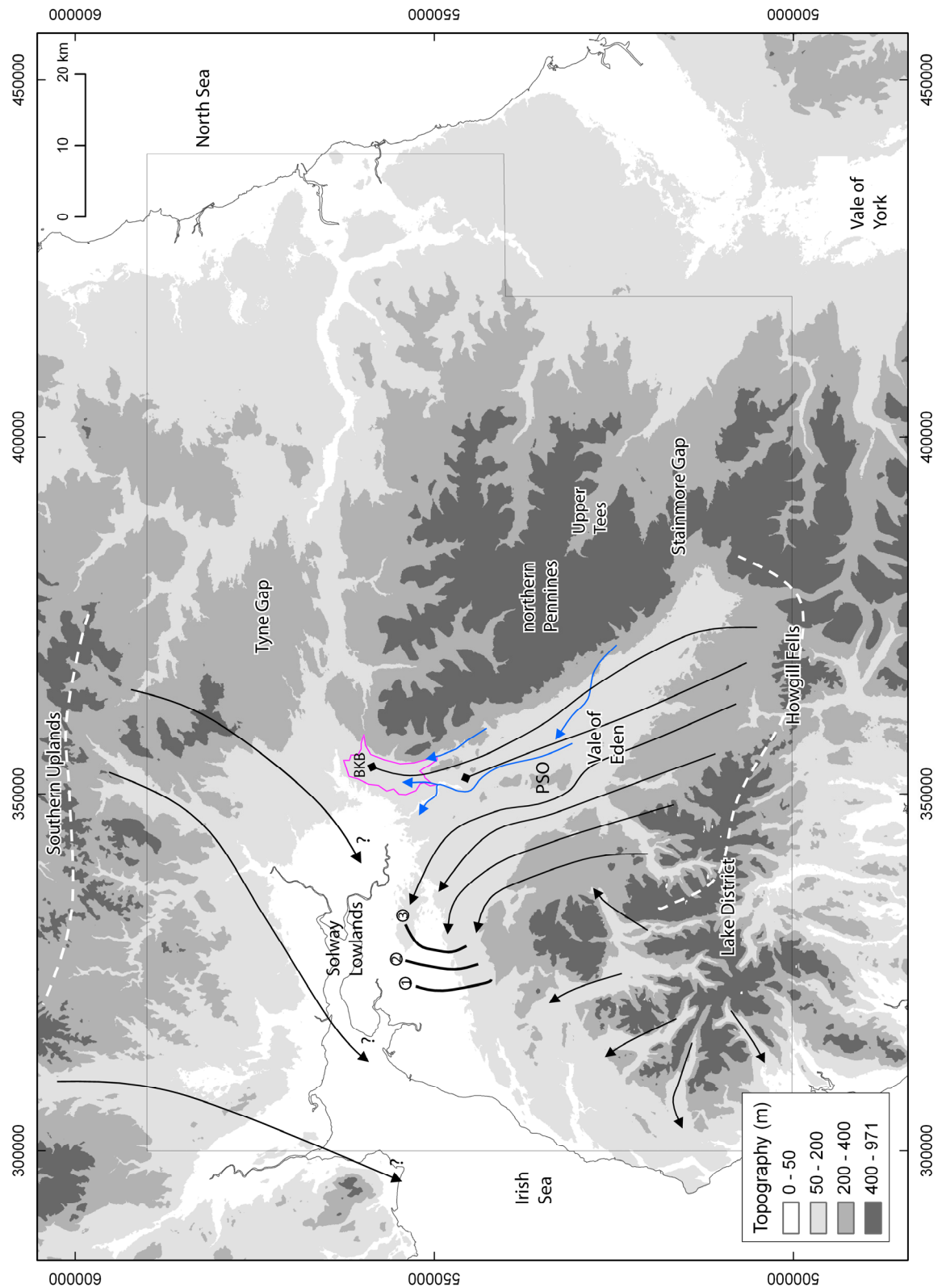
#### 9.2.6 Stage V

##### Retreat of ice from the Solway Lowlands:

Following the Blackhall Wood Re-advance ice continued to retreat out of the lowlands of the central sector of the BIIS (Fig. 9.5). Initially ice receded clockwise, around the northern margin of the Lake District, and NE back into the Scottish Southern Uplands (Fig. 9.5a). Local glacial-flow out of the northern Lake District became decoupled as the ice retreated eastwards, with ice-marginal or subglacial meltwater channels which terminate in perched fans delimiting successive still-stand positions (Hollingworth, 1931; Eastwood, *et al.*, 1968; Chapter 7; Table 9.1); similarly ice-marginal channels constrain the retreat of Scottish ice (Chapter 7; Fig. 9.5). Down-wasting at the northern boundary of the arcuate ice flow resulted in a considerable flux of meltwater, as recorded by glaciofluvial and deltaic deposits at Carleton (Chapter 7). Such an environment is replicated throughout the northern sector of the Solway Lowlands, with tripartite divisions clearly demonstrating the re-advance of Scottish ice (Stage VI) over glaciofluvial/glaciolacustrine deposits (Dixon *et al.*, 1926; Trotter, 1929; Phillips *et al.*, 2007; Stone *et al.*, in press; Chapter 7). During this stage Glacial-Lake Carlisle formed against the reverse slope of the Tyne Gap (Trotter & Hollingworth, 1932; Huddart, 1970; Chapter 7; Fig. 9.5; Table 9.1). The lake was impounded against ice in the Carlisle region and characterised by falling water levels and westerly expansion in response to further retreat and down-wasting of the two ice lobes (Fig. 9.5b; Chapter 7). The temporal evolution of the lake margin is constrained by a series of deltas, formed between 60 and 43 m O.D. (Huddart, 1970; Livingstone *et al.*, 2008). It must be noted that Glacial-Lake Carlisle might also/alternatively have formed during stage III, as ice retreated westwards away from the Tyne Gap. The lack of sedimentological evidence related to the Blackhall Wood Re-advance in this region makes this difficult to discern, although as the glaciofluvial and glaciolacustrine deposits are positioned stratigraphically below the Scottish Re-advance diamicton they have been correlated



with stage V.



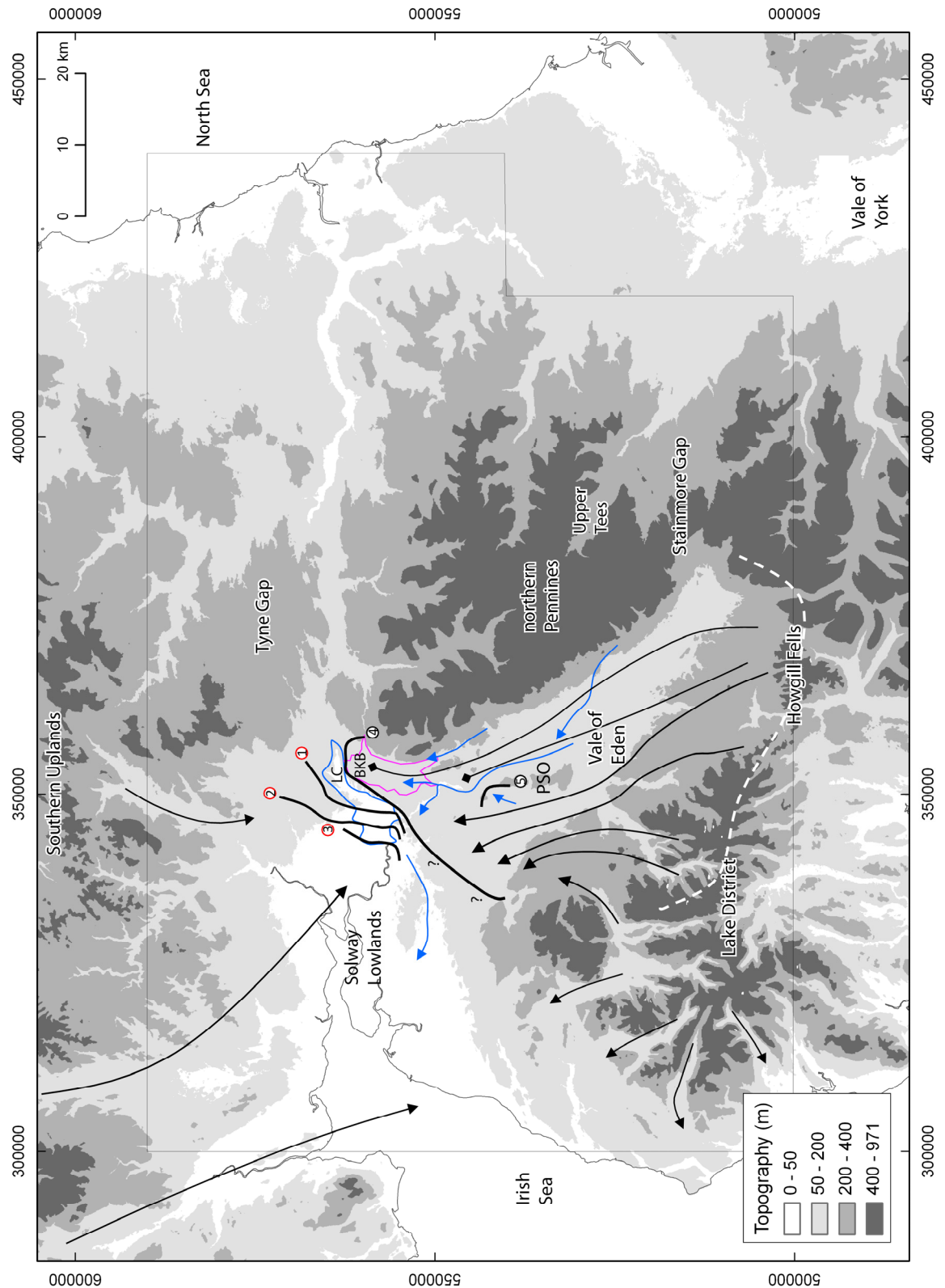


Figure 9.5: Stage V: Retreat of ice out of the Solway Lowlands. PSO stands for Penrith sandstone outcrop. This stage involves a complex series of glaciofluvial and glaciolacustrine deposition associated with the formation of the Brampton kame belt (BKB; pink polygon), Glacial-Lake Carlisle (LC; blue polygon) and the Pennine escarpment meltwater network (which formed time-transgressively throughout stages V-VI). 9.5a illustrates the initial retreat of both Scottish and Lake District ice out of the Irish Sea Basin. The black

<Fig. 9.5 (*continued*)... numbered circles refer to observed halt stages (1-3) (produced from perched fans) in the retreat of Lake District ice back towards the Vale of Eden. The position of the Scottish ice front at these halt stages are unknown. 9.5b refers to a later phase during the same retreat stage associated with the ‘unzipping’ of Lake District and Scottish ice. The westwards migration of the central Southern Uplands ice divide, coupled with topographical constraints resulted in a switch in Scottish ice flows, with the ice instead retreating westwards and northwards back into valley systems of the Southern uplands. The red numbered circles (1-3) refer to halt stages associated with the westwards expansion and associated fall in water level of Glacial-Lake Carlisle (produced from deltaic deposits). The black numbered circles (4-5) refer to further retreat stages of the Lake District ice as it continued to retreat back into the Vale of Eden and stagnate within the Penrith sandstone outcrop.

As with Glacial-Lake Carlisle, the Brampton kame belt (Livingstone *et al.*, 2008, in press c) could conceivably have formed during stage III or V (or throughout stages III-V). Given that the kame belt hugs the NW edge of the northern Pennines and is also associated with a suite of meltwater channels extending along the Pennine escarpment and into the Tyne Gap (cf. Livingstone *et al.*, 2008, in press c), it is logical to assume that this landform-assemblage can be correlated with the retreat of ice from the Tyne Gap during stage III. However, the meltwater channels along the Pennine escarpment are consistent with the NW direction of the youngest subglacial lineations in the region and therefore record the SE recession of the final ice to occupy the Vale of Eden (stages V and VI) (Evans *et al.*, 2009). It is therefore envisaged that the Brampton kame belt and associated meltwater channels formed in a time-transgressive manner (i.e. through stages III-VI; Table 9.1), with its physiographic position in the lee of the Pennines and pinned against the Penrith sandstone ridge facilitating stagnation and *in situ* downwasting of the ice mass (Livingstone *et al.*, in press c). Initial recession, westwards across the Tyne Gap (stage III), was characterised by subglacial meltwater breaching the Irthing-Tyne watershed (e.g. Gilsland meltwater channel and subglacial esker development) and feeding into the South Tyne Valley proglacial drainage network (Livingstone *et al.*, in press, in press c; Fig. 9.3). Further retreat into the Vale of Eden during the formation of ice-free conditions in parts of the Solway Lowlands corresponded to ice stagnation in the Brampton district (stage V, Fig. 9.4). Because the Brampton region is marginal to the Blackhall Wood re-advance and protected by the Penrith sandstone ridge it is reasonable to assume that initial formation survived remoulding. Following the ‘Blackhall Wood’ Re-advance ice in the Brampton district is thought to have evolved into a complex ‘glaciofluvial moraine’ with sedimentation controlled primarily by an enlarging glacier karst (cf. Livingstone *et al.*, in press c). The spatial and temporal evolution of the Brampton kame belt resulted in an array of landforms and sediment assemblages comprising flat-topped hills, depressions and ridges (Livingstone *et al.*, in press c). These are interpreted as ice-walled lake plains, ice-contact meltwater drainage networks and kettle holes respectively (Livingstone *et al.*, in press c).

The Brampton kame belt was fed by a large meltwater channel network trending along the edge of the Pennine escarpment (Totter, 1929; Arthurton & Wadge, 1981; Greenwood *et al.*, 2007;

Livingstone *et al.*, 2008, in press c), and also from meltwater draining off the Penrith sandstone ridge (Livingstone *et al.*, 2008, in press c). The drainage network documents the progressive deglaciation and down-wasting of ice in the Vale of Eden (stages V and VI) and comprises anastomosing subglacial channels and flights of lateral channels (Fig. 9.5). The morphology of the meltwater system as a whole is best explained as time-transgressive with channels running parallel to each other and formed at successively lower elevations (Livingstone *et al.*, 2008). This eventually cut-off meltwater drainage into the Brampton kame belt (Fig. 9.5). It is also tentatively suggested that the current River Eden Valley was occupied by a large tunnel valley during the last glaciation (Livingstone *et al.*, in press c). As ice retreated up the Vale of Eden during stage V, and following the end of stage VI, the active margin stagnated within the Penrith sandstone ridge (Fig. 9.5) forming a series of fans, deltas (e.g. Baronwood complex: Huddart, 1970) and corresponding meltwater channels (Livingstone *et al.*, 2008). Further meltwater channels have been mapped at the southern end of the Vale of Eden (Livingstone *et al.*, 2008), comprising four distinct dendritic networks, each of which leads into a major trunk channel. (Letzer, 1978; Livingstone *et al.*, 2008). These meltwater systems trend SSW-NNE, running parallel to the orientation of the drumlins in the region (stage III-V) and are interpreted to be subglacially formed due to their complex morphology and discontinuous profiles (cf. Letzer, 1978).

### 9.2.7 Stage VI

#### Scottish Re-advance:

The final stage in the observed glacial history of the BIIS is the re-advance of Scottish ice into the Solway Lowlands (ice-flow phase SF1: Chapter 2) (Chapter 7; Fig. 9.6; Table 9.1). This SE movement of Scottish ice across the Solway Firth and onto the fringe of the west Cumbrian coast impeded meltwater drainage, resulting in the formation of a large ice-dammed lake in the vicinity of Wigton (Chapter 7; Fig. 9.6). The eastern margin of the ice front was bounded by an ice-contact delta, and foreset structures reveal the lake height to have been 49 m O.D. (Chapter 7). Subdued eskers and evidence of a discontinuous thin upper till suggests a further transient advance inland (Chapter 7; Fig. 9.6). Lobate re-advance of Lake District ice is evident from a S-N orientated flowset of subglacial lineations encroaching into the Solway Lowlands from the Vale of Eden, although whether or not it occurred synchronously is difficult to discern (Livingstone *et al.*, 2008; Chapter 7; Fig. 9.6). This re-advance is bounded at its northern end by a belt of thick diamicton interpreted as an end-moraine (Fig. 9.6), while meltwater channels emanating from the former glacier margin, and coalescing with the Dalston overspill channel, are also evident at the northern edge of the Vale of Eden (Chapter 7; Fig. 9.6). As ice down-wasted and retreated into upland massifs flow became topographically constrained as depicted by a series of final flows which are clearly restricted to valleys of the main upland dispersal centres (ice-flow phase LT7: Chapter 2).

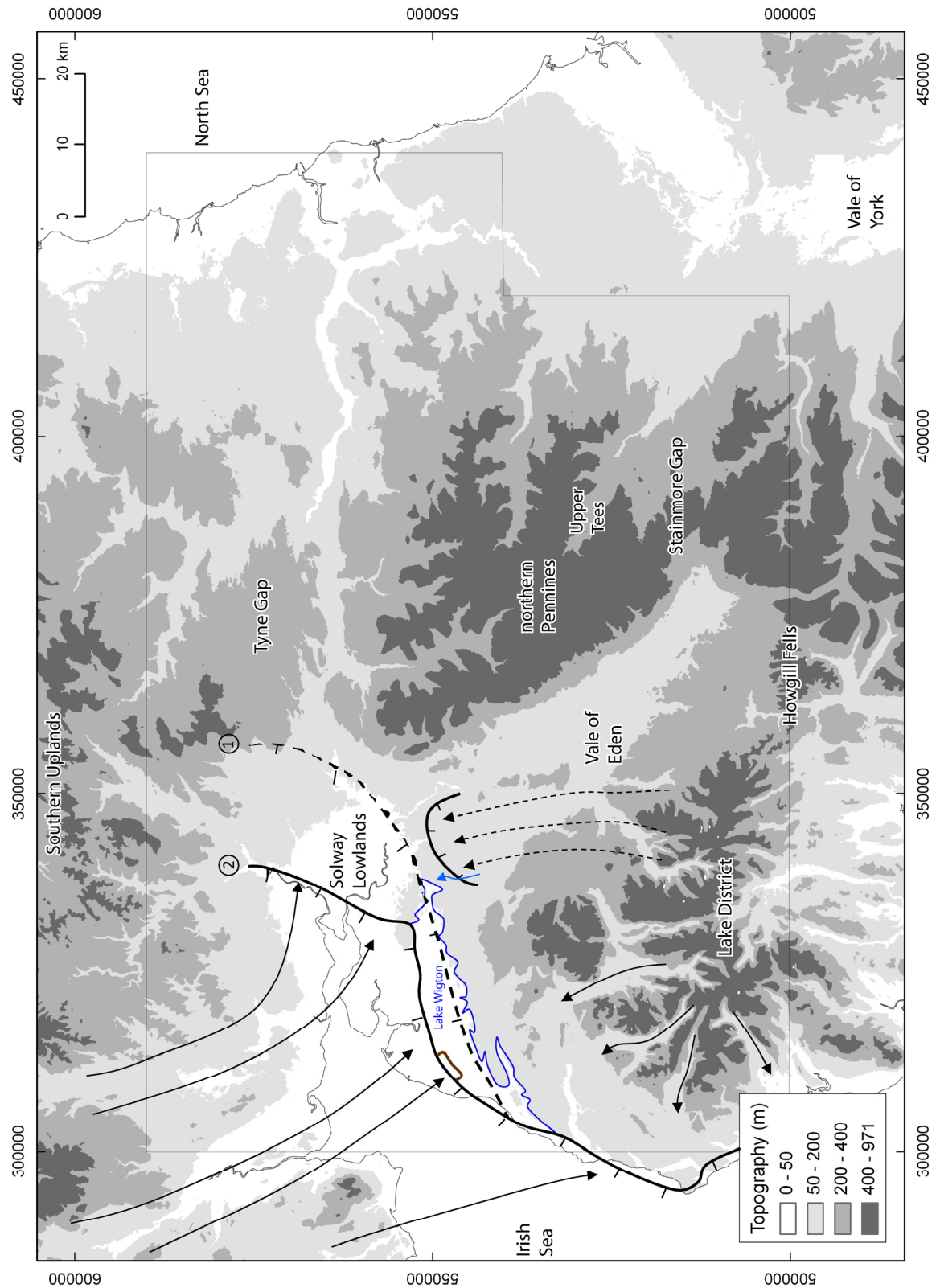


Figure 9.6: Stage VI: Scottish Re-advance. The brown box marks the location of Holme St. Cuthbert deltaic complex, while the dotted black line (number 1) refers to a transient (possible surge) of ice inland. The dotted arrows marking the position of the Lake District Re-advance demonstrate the uncertainty associated with assigning both re-advances to the same stage. The extent of Glacial-Lake Wigton, which formed during halt stage 2, is demarcated by the blue outline.

### 9.3 Conclusions

This research demonstrates the inherent dynamism of the central sector of the BIIS during the last glaciation. The ice sheet was characterised by flow switches, initiation (and termination) of ice streams, draw-down of ice into ice streams, repeated ice-marginal fluctuations and the production of large volumes of meltwater, often impounded to form ice-dammed glacial lakes. The topographic complexity of the region provides an underlying control for the multi-phase ice flow patterns exhibited. Upland regions provided both the initial and final focal points from which ice radiated (under topographic control), while during later phases ice was streamed through topographic lows (e.g. Tyne Gap) and influenced by changes in the dominance of upland IDCs. Dynamic shifts in flow direction were primarily driven by the migration of IDCs and ice divides. This is depicted in the generalised migration patterns of ice divides and IDCs of the BIIS, with ice first expanding and then contracting back into upland massifs (e.g. Evans *et al.*, 2009). Ice streams both within and in the immediate vicinity of the central sector of the BIIS also influenced the dynamics of the ice sheet through the draw-down (and thus thinning) of the ice (e.g. Irish Sea Ice Stream). This impacted upon the migration of ice divides initiating rapid changes in flow behaviour and direction; e.g. the Tyne Gap. Ice streams themselves are observed to be non-permanent and highly sensitive features of the BIIS; with the Tyne Gap, for example, shown to be heavily influenced by both the Southern Uplands and Lake District IDCs, before eventually disintegrating as the Irish Sea ice divide shifted northwards. The Irish Sea Ice Stream however, did not influence the central sector of the BIIS until a late stage of deglaciation, probably with the ice divide in the northern part of the Irish Sea Basin instead providing the major flux of ice from this sector. Ice flow shifts are shown to be common-place in what is a highly mobile and dynamic region of the ice sheet. Numerical ice-sheet modelling demonstrates that these ice-flow switches can occur rapidly over short time-scales. This depiction of a dynamic ice sheet heavily influenced by fast-flowing ice streams is in accord with recent fieldwork and modelling studies that recognise dynamic shifts in ice flow direction during the glaciation of the last BIIS (e.g. Greenwood & Clark, 2008; Hubbard *et al.*, 2009).

Deglaciation was both complex and non-linear, with numerous oscillations and switches in ice flow, similar to other regions in the Irish Sea Basin. Two major oscillations (the Blackhall Wood Re-advance and Scottish Re-advance) have been identified within the central sector of the BIIS. Both re-advances are thought to be part of more regionally extensive Irish Sea Ice Basin re-adjustments and can be tentatively correlated to the Gosforth Oscillation at ~19.5 cal. ka BP (Merritt & Auton, 2000) and Killard Point Stadial at ~16.8 cal. ka BP (McCabe *et al.*, 2007) respectively. These re-advances can be attributed to external forcings related to the ~19 ka cal. BP meltwater pulse (cf. Clark *et al.*, 2004) and ~16.8 cal. ka BP Heinrich Event-1 (c.f. McCabe *et al.*, 2005) (Table 9.1). Throughout the period of deglaciation this sector of the ice sheet was characterised by the production of large volumes of meltwater, leading to deeply incised ice-



marginal, pro- and sub-glacial channels. Indeed, the repeated and widespread development of ice-contact lakes (e.g. Stokes & Clark, 2004) not only in the Solway Lowlands, but throughout the NE sector of the Irish Sea Basin, provides a possible internal mechanism for the de-stabilisation of the Irish Sea Ice Stream through rapid drainage of large volumes of meltwater. This type of glacial environment also has implications for the movement of ice in the Solway Lowlands (Stokes & Clark, 2004), with similar environments in SW Scotland thought to promote surging of the ice margin (Salt & Evans, 2004). The temporal evolution of ice retreat was non-linear with stagnation and *in situ* down-wasting in the Brampton region (kame belt), at the eastern edge of the Tyne Gap and also along the Penrith sandstone ridge. Elsewhere, moraines and ice-contact deltas demarcate still-stands associated with ice-frontal retreat, while meltwater channels document the style of ice wastage.

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# Glacial geomorphology of the central sector of the last British-Irish Ice Sheet

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Mapping based on NEXTMap data (NEXTMap Britain data from Intermap Technologies Inc. were provided courtesy of NERC via the NERC Earth Observation Data Centre).  
British National Grid reference system. Horizontal Datum: ETRS89. Vertical Datum: ODN (OSGM91).  
Projection: Transverse Mercator, airy spheroid, OSGB36.

Legend:

Rivers

Water bodies

Glacial Features

Glacial sands and gravels

Ribbed Moraine

Hummocky Terrain

Meltwater channels

Eskers

Lineations, colour coded in relation to interpreted ice flow phases

A Dumfries-shire

Phase 1

B1 Tyne Gap

Phase 1

Phase 2

Phase 3

Phase 4

Phase 5

Phase 6

Phase 7

Phase 4-7

Phase 1-7

B2 Tyne Gap

Phase 1/2

Phase 1/2

C Vale of Eden

Phase 1

Phase 2

Phase 3

Phase 4

Phase 5

Phase 1-2

Phase 1-4

Phase 1-5

D1 Stainmore-Howgill Fells

Phase 1

Phase 2

Phase 3

D2 Stainmore-Howgill Fells

Phase 1

D3 Stainmore-Upper Tees

Phase 1

Phase 2

Phase 3

D4 Stainmore

Phase 1

Phase 2

Phase 3

Phase 2-3

Phase 1-3

E East Coast

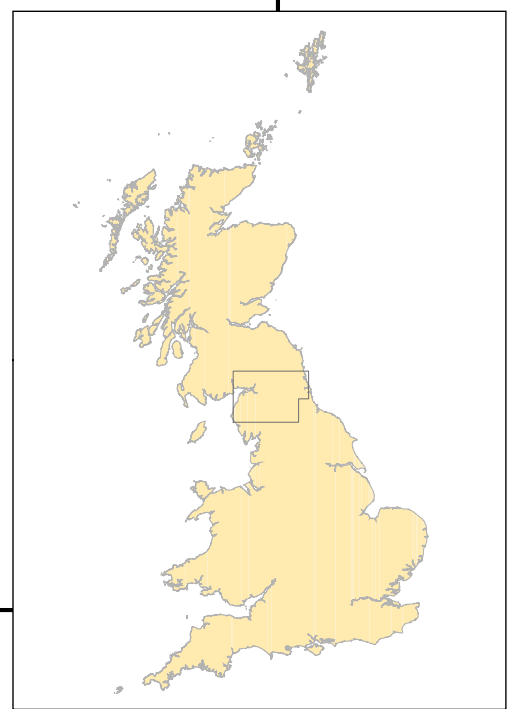
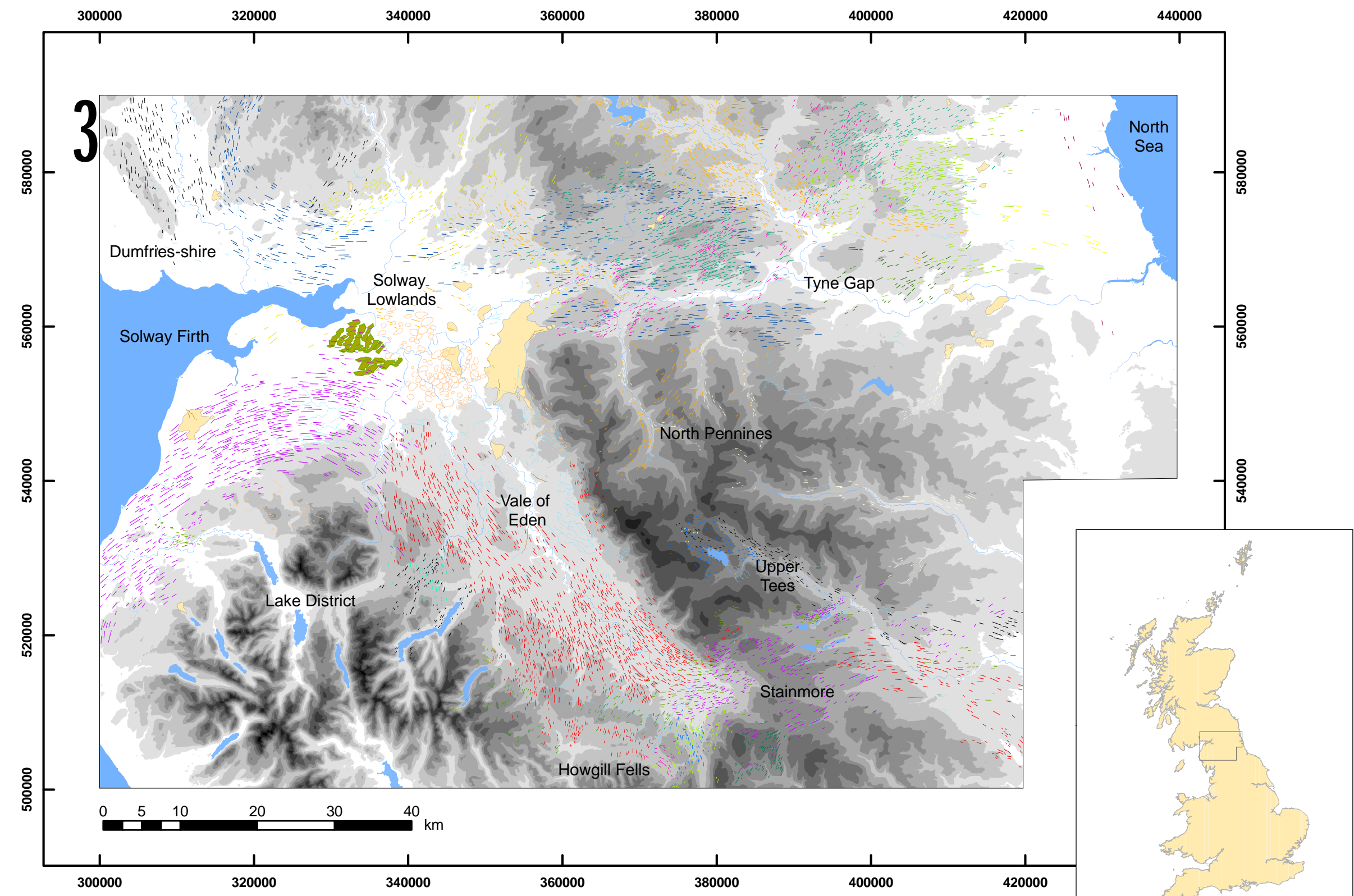
Phase 1

F North Pennines

Phase 1

G Valley Glaciers

Pennines



Spatial Resolution: 5m posts  
Vertical Accuracy: 1m